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SOME EVENTS IN THE EOGENE HISTORY OF THE PRESENT COASTAL AREA OF THE GULF OF MEXICO IN TEXAS AND MEXICO

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EOGENE SECTION

The most complete section of the Eogene[1] deposits of the eastern border of North America is found in Alabama and Mississippi, but even this is somewhat indefinite in its upper members

[1] Used as inclusive of both Eocene and Oligocene.

and only by supplementing these with their marine representatives of the Florida region can the entire section be clearly understood

Thus constructed, the section would be:

	Alabama–Mississippi	Florida
Oligocene:	Grand Gulf	Tampa
	Ocala	Chattahoochee
	Vicksburg	
Eocene:	Jackson	
	Claiborne	
	Wilcox	
	Midway	

The deposits as a whole are those of marine waters alternating with estuarine and fresh-water conditions, and the fossils indicate uniformly a warm or subtropical temperature of water.[1]

The time relations of the various members of these groups can be best made out in such an area of apparently continuous sedimentation as that found in this typical area, but when, as in this case, a study of the faunas of the various members indicate great physical changes intervening between them, further evidence of such changes is to be looked for in the extensions of the deposits. Some important evidence of this character is to be found in the Texas-Mexican region (Fig. 1).

CRETACEOUS-TERTIARY BARRIER

Prior to the deposition of the earliest Tertiary sediments in this area the Cretaceous rocks had been subjected to vulcanism, folding, and erosion. These movements began during the period of the Austin Chalk, at which time a barrier of some sort was created in the vicinity of the Salado and Sabinas rivers in northeastern Mexico which formed the beginning of what has been called the Rio Grande embayment. To the north of this barrier the succeeding Cretaceous sediments are largely clays and sands and include the coal beds of the Rio Grande region, while to the south we find only a great thickness of blue and black shales. The beds to the north are frequently very fossiliferous, while the shale beds are, so far as known, practically destitute of any trace of fossils.

[1] Dall, *Trans. Wag. Inst.*, p. 1549.

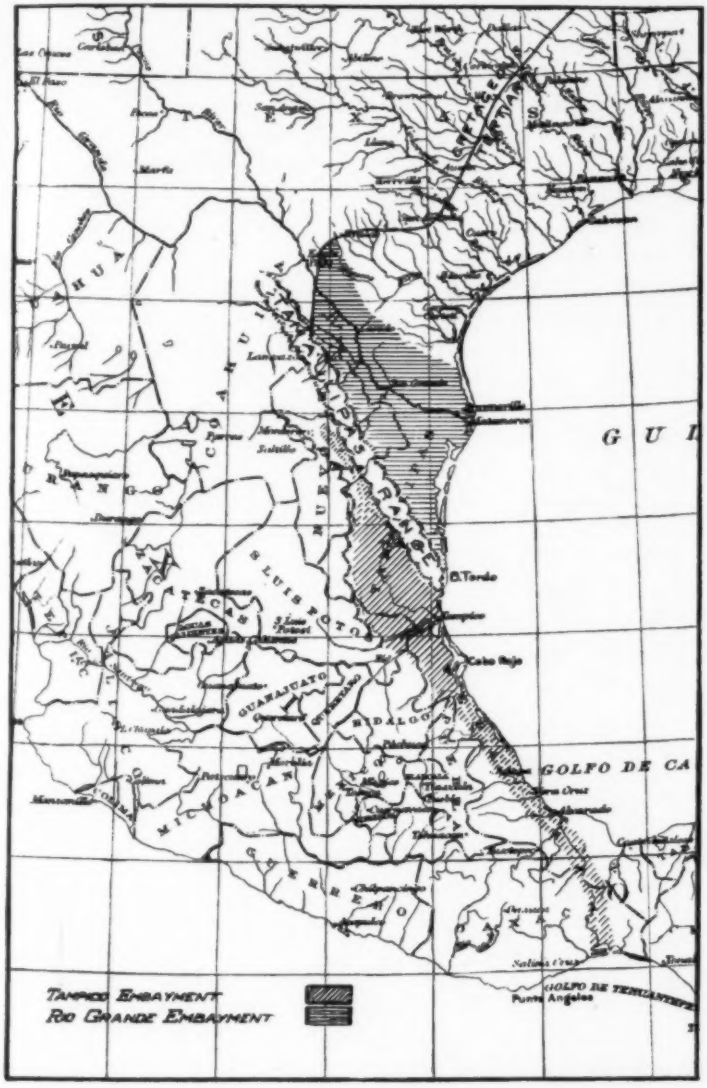

FIG. 1. Barrier in Gulf Coast region separating Atlantic and Pacific waters in Cretaceous-Tertiary times.

This movement was marked in the Texas area by Pilot Knob and other volcanoes south of Austin, which were active during the close of the Chalk deposition and the beginning of the Taylor marls, as is shown by the ash from them, which is interstratified with, and included in, these deposits.

At the close of the Cretaceous the movement was intensified and resulted in a land barrier which is now marked by the disconnected ranges and groups of hills that form the eastern border of the valley lying at the foot of the Mexican Cordilleras.

These groups and ranges include the San Antonio, San Juan, Vallecillo, Picachos, Papagallos, San Carlos, and Tamaulipas, for which, as a whole, Professor Cummins has proposed the name of the Tamaulipas range. Since the trend of the coast in this region is a little west of south, the southeast course of this Tamaulipas Range brings it rapidly nearer the Gulf. At Tordo Bay, some fifty miles north of Tampico, the main body of the hills is within ten miles of the coast, while outliers extend almost to the water's edge.

This range is made up of shales and limestones, more or less altered and disturbed by folding and by igneous rocks. No fossils have been found in the shales, but the limestones yielded a few *inocerami* and fragments of *Ammonites*. Dr. Stanton, who examined these, states that these limestones are not younger than the Taylor marls and may not be younger than the Austin Chalk.

The Tamaulipas Range represents the extension southward and culmination of the movement creating the Sabinas barrier, and just as the Sabinas barrier forms the southern limit and border of the Gulf Coast Cretaceous deposits, so does the Tamaulipas Range form the southern border and limit of the Gulf Coast Eocene. The difference in the character of deposits laid down in late Cretaceous time to the north and south of the Sabinas barrier is just as strongly shown in the difference between the Eocene deposits east of the Tamaulipas Range and those of the same age which are found to the west and south of it, since not only are the deposits themselves of different character, but even the fossils which the latter carry are entirely unlike those to the north.

CRETACEOUS-TERTIARY CONTACT

South of the Salado River the exposures of the Cretaceous shales along the line of contact with the Tertiary indicate that they have been subjected to more or less folding and to erosion prior to the incursion of the waters of the Eocene, so that the contacts from Rodriguez south show very decided unconformities.

On the Rio Grande the unconformity is shown partly by difference in direction of dip and partly by the fact that the Midway is found resting upon different members of the Escondido.

Northward from the Rio Grande the uppermost beds of the Cretaceous present, as a whole, a gentle slope toward the Gulf, and when we find contacts between these beds and the basal Tertiary the discordance of stratification is usually very small, even if it be present at all. There is, nevertheless, ample proof of folding in this area also, for logs of wells drilled in the Tertiary belt of western Louisiana and eastern Texas give positive evidence of broad folds in the underlying Cretaceous beds far to the seaward of their present outcrop.

The most striking feature of the orogenic movement in this area is, however, found in the numerous submerged hills of the salt domes that now occur as inliers in the Tertiary belt. These, as shown by the discordance of dip between the Cretaceous rocks and those of the surrounding Eocene, are pre-Tertiary.

LOWER EOCENE

Midway.—The initial sediments of the Eocene were marine and are fairly uniform in character throughout their entire extent within the area of the western Gulf Coast. While east of the Brazos they occupy a belt which has at times a width of several miles, with a total thickness of between 300 and 400 feet, west of that stream they are found in disconnected exposures due generally to the overlapping of later deposits.

This is clearly shown on the Rio Grande, where excellent contacts have been found by Kennedy, and where this substage attains a thickness of 300 feet. Here the Midway is frequently covered both by the Wilcox or Lignitic and by the Carrizo sands or basal Claiborne.

These conditions continue to the Salado River, but south of that stream the Midway covers a much larger surficial area and the Wilcox and Carrizo are less prominent.

By far the best exposures of the Midway which we have found are those in the valley of the Salinas River east of Ramones, Mexico. Here they occupy a considerable area and also form the Alto Colorado, a hill 250 feet in height which is composed entirely of Midway sediments. These have a total thickness of more than 400 feet. Among the fossils found in some abundance are *Ostrea pulaskensis*, *O. crenulimargo*, *Venericardia alticosta*, *V. planicosta*.

The last exposure of these beds within this area was found on the road to China, six miles southeast of Ramones.

That the close of the Midway was brought about or accompanied by an uplift is proven in the Rio Grande region, where it is seen to have a greater dip than the succeeding Lignitic series.

Wilcox or Lignitic series.—The series of fresh-water or estuarine deposits which followed the Midway shows little change in the general character of its sediments southwestward, with the exception that south of the Colorado it loses the larger part of the lignitic beds which are such a prominent feature of it north and east of that river.

North and east of the Colorado the Wilcox covers a broad area and has a thickness, as shown by wells drilled through it, of 1,000–1,200 feet. South of this river the exposure gradually becomes narrower and the beds thinner.

Its exposures along the Rio Grande are few and the thickness of the beds exposed probably does not exceed 100 feet. All of its observed contacts with the Midway were unconformable and there is a distinct difference in the degree of dip of the beds of the two stages. The sand and clays of the Wilcox Lignitic show evidence of emergence and erosion and in many places are entirely absent, having evidently been eroded prior to the deposition of the lowest Claiborne sands.

The Lignitic is found within a few miles south of the Rio Grande in a number of exposures. Farther south and beyond the Salinas it overlaps and covers the Midway, so that the latter

is seen no more and the Lignitic is itself in contact with the shales of the Cretaceous.

The most southerly exposure of the Lignitic was seen on the Conchos River a mile west of Angeles, where it is uncovered for a short distance by the erosion of that stream.

At the close of the Lignitic its sediments emerged and were strongly eroded prior to the incursion of the Claiborne sea.

In Lower Eocene time, therefore, the area north of the Tamaulipas Range was submerged and marine waters prevailed as far south as the latitude of Monterey. At the close of this period the marine waters receded and after a period of erosion were followed by the submergence of the deposits under fresh or brackish waters. The close of this second or brackish-water period of sedimentation was followed by its emergence and erosion.

In the region west and south of the Tamaulipas Range no deposits have been found which can certainly be referred to the Lower Eocene as represented north of that range by the Midway and Wilcox.

Beginning near the foot of the high plains of the interior of Mexico there is a series of blue and gray shales which have a wide development, forming a large part of the surface rock of the valley between the Cordilleras and the Tamaulipas Range and continuing southward beyond the Tuxpam River as far as examinations were carried. In the northern part of this area these shales have been traced to a direct connection with those found in the San Juan and Papagallos mountains, which, by their relations to the basal Tertiary and the fossils found in limestones immediately underlying them, were proved to be Upper Cretaceous.

There is undoubtedly a great thickness of these shales, but no fossils have ever been found in them. There is, therefore, a possibility that in their upper portion some representative of early Eocene time may be present, but, if so, it has not yet been differentiated.

So far as yet recorded no fossils characteristic of the Midway or Lignitic have been found between Tordo Bay and Yucatan, although fossils of both periods are known to exist in Venezuela and elsewhere to the south.

MIDDLE EOCENE

Buhrstone and Claiborne.—In the Alabama section the Buhrstone and Claiborne have a total thickness of about 450 feet, the Claiborne proper being only one-third of this amount. The surface exposures rarely exceed a total width of fifteen miles. In Texas the narrowest portion of this Claiborne belt occurs on the Colorado River and has a width of twenty-five miles, which widens northeast to ninety miles on the Neches and attains a similar width southwestward in the Rio Grande region. Along the Rio Grande, which crosses the formation obliquely to the dip, the belt has an exposure of 150 miles. The beds in Texas also have a correspondingly increased thickness and have been divided as follows: the Carrizo sands corresponding to the Buhrstone; the Marine, Yegua, Fayette, and Frio substages—all of which carry fossils characteristic of the Lower Claiborne. These substages are present from the Conchos or Presas River in Mexico to the Colorado in Texas. The Frio has not been recognized east of the Colorado, and east of the Neches the Fayette is also wanting, except as isolated patches, owing to pre-Jackson erosion.

The Carrizo sands which mark the beginning of the Middle Eocene deposition are the stratigraphic continuation of the Queen City beds of northeast Texas. Throughout this region where Kennedy first observed them, the relations of these sands to the underlying Lignitic is such that he considered them to be the upper member of that stage and so described them. Later examinations in the western part of the state, however, brought out the fact that the Carrizo sands were entirely unconformable with the Lignitic deposits and that in some localities they grade upward into the basal beds of the Marine substage.

In Northeastern Texas the Carrizo sands have a thickness of 70 feet and appear to be conformable with the Lignitic. On the Atascosa the contact was not observed, but the sands were apparently of much greater thickness. On the Rio Grande the beds are exposed over a very wide area, outcropping along the river for a distance of forty miles. They have a less marked dip than either the Lignitic or Midway and are found resting on both of these and

overlapping them to contacts with the Escondido beds and the Papagallos shales of the Cretaceous.

South of the Rio Grande the Carrizo sands are well exposed and form the crest and eastern slope of various small groups and ranges of hills lying between the Rio Grande and Azulejo and of the larger range south of that point, known as the Ceja Madre. These hills have a total length of about ninety miles and reach nearly to the Salado River.

South of the Salado the Carrizo is not so prominent and was observed only in a few localities.

The Marine beds in places, as on the Rio Grande and the Atascosa, are directly connected with the Carrizo sands by transitional beds of lignitic shales, but at other places, and especially in east Texas, no such connection can be found and beds of greensand or of brown sandstone are directly superposed upon the Carrizo, although usually without apparent unconformity. Local movements during Marine time are, however, registered in the sediments in some parts of the area.

The Marine beds have a thickness of 800–1,000 feet. The eastern portion of the deposits seems to have been most highly glauconitic, although the beds everywhere give evidence of the presence of this mineral in greater or less quantity.

In the eastern portion of the state the alteration of this glauconite has resulted in the formation of extensive deposits of excellent iron ores, but in the western portion the resulting ferruginous matter is widely disseminated and simply forms a cement for the brown ferruginous sandstone which covers such a great area in Southwestern Texas.

On the Rio Grande the carbonaceous shales of the Carrizo find their culmination in the San Pedro and San Tomas beds of cannel coal between which are beds bearing characteristic Claiborne fossils.

The exposure of the Marine beds on the Rio Grande has a width of more than fifty miles. They narrow rapidly to the south and where they cross the Salado west of Guerrero the outcrop is not more than five or six miles wide.

They are well shown in the valley of the San Juan and are traceable for twenty-five miles south of that stream, when they are covered by the Yegua. They are not observed on the Conchos.

Fossils are abundant throughout the entire extent of the substage, although the upper portion is usually the most fossiliferous portion. In the Texas area in addition to the characteristic fauna we find a few forms practically identical with those of the Tejon. In the San Juan region, associated with the *Venericardia planicosta*, there are found vast numbers of a varietal form very similar to, if not identical with, *V. potopacoenses* of the Maryland Eocene.

The succeeding substage, the Yegua clays, appears to be unconformable with the Marine. The Yegua deposits are predominatingly lignitic clays, but comprise beds of sands and greensands as well. They, unlike the earlier Lignitic, maintain their thickness of 1,000–1,400 feet and carry their deposits of lignite entirely across the state. They are also the principal gas-bearing beds of the Eocene. Like the lower Lignitic, they become more sandy on the Rio Grande and this condition persists throughout exposures south of the river to Mendez on the Conchos or Presas.

In their eastern part these beds are not as fossiliferous as they are west of the Guadalupe, although they carry fossils in many places, especially between the Brazos and the Colorado. On the Rio Grande they are nearly as fossiliferous as the Marine.

It has been suggested that the so-called Cocksfield Ferry beds on the Sabine might be of Upper Claiborne age, but they have been connected by careful tracing with the basal Yegua and are, therefore, Lower Claiborne.

To these deposits of Yegua clay there succeeded another substage, the Fayette, which, while it carries beds of lignitic clays and sands with some lignite, is principally composed of sands and sandstones. Some of the sands are friable while others are quartzitic and they carry a considerable amount of fossil wood which is opalized instead of being simply silicified. They have a thickness of 500–600 feet.

East of the Guadalupe, with rare exception, the Fayette does not carry any fossils except the wood, but west of that stream it contains the large oyster, *O. alabamiensis var. contracta* of Conrad,

which is one of its most characteristic forms. On the Nueces and Rio Grande we find, besides this oyster, numerous other forms which link it positively with the Claiborne. This is especially true of the exposures on the Rio Grande which have a width of fully forty miles with many fossil beds. These include:

Venericardia planicosta, Lam.
Anomia ephippioides, Gabb.
Leda opulenta, Con.
Crassatella protexta, Con.
Tellina mooreana, Gabb.
Cytherea bastropensis, Har.
Conus sauridens, Con.
Volutilithes petrosus, Con.
Levifusus trabeatoides, Har.
Lacinia alveata, Con.
Pseudoliva vetusta, Con.
Turritella nasuta, Gabb.

South of the Rio Grande it is found in a broad belt extending southeastward by way of Mier, Aldamos, China, and Mendez to the Conchos and beyond as far as the Chorreras, where it finally passes under cover of the Oligocene.

Between the Conchos and the Chorreras the Fayette apparently forms a portion of Mt. Corcovada east of Burgos, being elevated by a basaltic extrusion.

The Frio clays which follow are found overlying the Fayette sands from the Conchos River to the Guadalupe. On the Conchos and on the Ramones Mountains lying north of it they show a considerable thickness of yellow and darker clays which weather white and are accompanied by considerable quantities of disseminated gypsum and also beds of this material. Farther north the Frio beds are not so thick and in many places only a few feet of the formation are found between the sands of the Fayette and the overlying Oakville sands of Miocene age. Such fossils as have been found are apparently of Claiborne age and the clays are, therefore, referred to that stage.

The Middle Eocene of this Gulf area represents a succession of alternating marine, lagunal, and swamp conditions. The marine conditions were most constant in the Rio Grande region and farther

south, while in the eastern portion of the state the lagunal and swamp conditions apparently predominated.

There are unconformities here and there and evidences of local warping of the strata in the various substages, but they are, nevertheless, bound together by transitional beds or identity of fossils.

The constantly changing conditions of deposition in the region east and north of the Tamaulipas range during Middle Eocene time, as shown by the variableness of the deposits above described, did not extend to the region lying west and south of that range.

West of the Tamaulipas Range the blue shale of the uppermost Cretaceous is followed by other shales so similar in general character that only the finding of Eocene fossils in them suggested the differentiation of the two.

The only localities where exposures of these beds have certainly been found is in the vicinity of Alazan, on the Buena Vista River, about twenty-five miles northwest of Tuxpam.

The deposits are bluish shales that weather white or yellow and their exact relations to the Cretaceous were not determined, on account of the inadequate exposures. The fossils are widely scattered through the shale and it is possible to collect them in good condition only from decomposed fragments of the rock.

This locality was discovered by Mr. DeGolyer and fossils found by him were recognized by Dr. Dall as Pacific Coast forms. It was afterward visited by Professor Cummins, who made a somewhat larger collection of the fossils. These were also submitted to Dr. Dall and afterward turned over to Professor Roy E. Dickerson for study in connection with other Pacific Coast faunas of Eocene age in his work at the University of California.

The collection of fossils made at this locality by Professor Cummins comprises a fauna consisting principally of gasteropods and small bivalves. The preliminary examination failed to show any forms which could be referred to our Claiborne, but did show some that were apparently identical with, and many that were very similar to, Eocene Tejon of California.

These beds dip eastwardly and probably extend some distance north and south of Alazan. Shales similar to these were observed along the railroad north of the Panuco River and on the San

Antonio River west of Cruz, a station on the railway between Tampico and Monterey. Cummins correlates these shales north of Panuco and west of Cruz with the Alazan deposits because of common lithological character, although he found no fossils in them.

They were, however, certainly found in drilling a well at Topila, on the Panuco River, at a depth of 1,810 feet, as shown by the fossils brought up with the drill. Of the fifteen species collected at this locality nearly one-half are identified as Tejon species and the others as nearly related to them.[1]

UPPER EOCENE

So far we have not positively recognized the Frio clays, which are the uppermost beds referable to the Middle Eocene, east of the Colorado. Other beds of clay have been referred to this substage, but closer investigation seems to show that the Frio clays are not present in eastern Texas. Growing out of such correlation it has been presumed that both the Fayette and Frio of eastern Texas are later than the Lower Claiborne and correspond with the Jackson and the Oligocene.

In the area west of the Colorado River, where these beds have had closest study and where they are most fossiliferous, it is very certain that this presumption is erroneous. There is a continuity of the Lower Claiborne fauna through these upper beds and no forms whatever of Jackson or later age have been observed.

In all this area the Frio is found resting seemingly upon an uneroded surface of Fayette sands and north of the Rio Grande it is unconformably overlain by the Oakville Miocene without beds of either Upper Eocene or Oligocene age intervening.

East of the Guadalupe River the conditions are somewhat different. In the valley of the Colorado and eastward to the Sabine it is evident that subsequent to the deposition of the Frio and prior to the beginning of Jackson deposition as now known, there was a period of erosion during which the Frio, more or less of the Fayette, and even a part of the Yegua were carried away. This erosion probably took place during the deposition of the Upper

[1] *Science*, XXXV, 906–8.

Claiborne sediments in other regions and appears not to have greatly affected that portion of our Claiborne belt lying between the Guadalupe in Texas and the Conchos in Mexico. During these oscillations of the land it is evident that the relief at any time was slight.

Jackson.—Beginning west of Groveton, in Trinity County, Texas, in the valley of the Neches, and stretching eastward to the Sabine is a belt eight or ten miles in width in which are found calcareous clays and sands with nodules and bands of impure limestone, carrying fossils of Jackson age. For the most part these deposits lie directly upon the clays of the Yegua, since only occasionally are the Fayette sands to be found between them.

The identity of these deposits is clearly evident from the considerable number of fossils found at various localities.

The Jackson sediments have not been found west of Groveton. In their place, however, the Fayette sands gradually assume their normal position in the section, showing less and less erosion, until the Colorado drainage is reached.

The Jackson shore line may have extended westward as far as the Brazos, for sands thought to be at the top of the Fayette near Wellborn have yielded fossils which are now said to be of Jackson age, although originally the collections were classed as Lower Claiborne.

These are supposed to be, however, a lower horizon in the Jackson than the deposits farther east.

In southeastern Texas we have found indications of the gulfward extension of the Jackson deposits in fossils from oil wells drilled at Saratoga and Sour Lake, but southwest of the Brazos no indications whatever of any materials of this age are known either in Texas or in Mexico.

We have, therefore, in this eastern Texas area simply the western limit of the Jackson deposition and the materials themselves in many places give evidence of the near-shore conditions.

Almost nothing has been found in this area that throws light on the happenings between the close of the Middle and Upper Eocene and the beginning of Oligocene deposition. Whatever there may have been is covered by the succeeding beds of the Oakville.

OLIGOCENE

Nowhere within the Texas area has any trace of Vicksburg strata been found, but Hilgard and Loughridge both contended that the Grand Gulf sandstones were found along the Sabine, and Loughridge traced them nearly across the state.

There is a belt of sandstone crossing the state just as described by this observer and it has some of the lithologic characteristics of the Grand Gulf. Our investigations seem to indicate, however, that while it may be traced in apparent continuity across the area it is in reality not homogenous, but is in fact a composite series. That it represents in part remnants of a deposit of the Grand Gulf sands, even if not fully susceptible of proof, is strongly probable. We also know that it comprises beds of later deposition which have surrounded and overlapped these Grand Gulf remnants. These later deposits are the Oakville sands of western Texas and are of Miocene age. Among the outcrops now remaining which we think are Grand Gulf are those at Colmesneil, the quarry north of Corrigan, near Riverside, etc.

The lack of fossils in the Grand Gulf sands and the close resemblance to them (even including the quartzitic phases) of those of Oakville age will probably prevent any complete differentiation of the two, although this may be done in part through a study of the scanty flora occurring in them here and there.

If any other Oligocene deposits besides the remnants of the Grand Gulf were laid down in the Texas area they have been eroded or are also covered by the overlap of the Oakville.

In Mexico, on the contrary, we find a broad development of the Marine facies of the Oligocene deposits such as are found in Florida and the Antilles as distinguished from the near shore brackish-water phase of most of the Gulf area.

North of the Tamaulipas Range we find, overlying the Eocene, a series of yellow sands, clays, and calcareous beds which carry an abundant Oligocene fauna. They were first studied in the vicinity of San Fernando, on the Conchos River, where there are fine exposures, and we have called this local development of the beds the San Fernando.

These Oligocene deposits are found resting upon the Frio clays along the Conchos River and the eastern front of the Pomeranes Range, and they extend northward to within fifty miles of the Rio Grande, beyond which point we find no trace of them, possibly because of the extensive development of the Reynosa surficial deposits in that region.

South of Abasola the Oligocene overlaps the Frio and is in direct contact with the Papagallos shales on the west and extends eastward to the Gulf shore.

The corals found in some of these beds have been studied by Dr. Vaughan, who says that they indicate an Upper Oligocene horizon, about equivalent to the Chattahoochee of Georgia. This is borne out by such of the molluscan remains as have been examined, and so far our collections from this area have not yielded any forms characteristic of the Lower Oligocene. Among the corals determined are: *Favites* (?) *polygonales* Duncan, *Goniastrea antiquaensis* Duncan, *Acropara* ? sp., *Orbicella cellulosa* Duncan, *Orbicella* n. sp., *Geneopora* sp. very similar to, or identical with, an Antiguan species.

South of the Tamaulipas Range the yellow clays and sands of the Oligocene cover a broad area. Between the Buena Vista River and the Tamiahua Lagoon many exposures are found and the beds are highly fossiliferous. *Cristalleria* and *Nummulites* abound and *Orbitoides papyracea* Bou. is found from the Buena Vista to the Tancochin at Cerro del Oro. Dr. Dall states that this form occurs in the Lower Oligocene beds in Alabama, indicating a Lower Oligocene age for this portion of the deposits. This Oligocene belt has a known width of twenty miles east of the Buena Vista River and has been observed twenty miles west of that stream resting upon the shales of the Cretaceous.

In addition to the fossils named there are many others, including several echinoderms, pectens, and other bivalves and numerous gasteropods.

While a detailed examination of these beds has not been made, enough has been done to prove that there is in this part of Mexico very considerable thickness of these Oligocene deposits and that they represent portions, if not all, of both Lower and Upper Oli-

gocene time. The lower beds, so far as now determined, are confined to the region south of the Panuco, while the upper stretch northward beyond the old Eocene barrier nearly to the Rio Grande.

MIOCENE

In Texas, as we have seen, such Oligocene deposits as may occur are overlain directly by the Miocene (Oakville) and the sands of one are almost indistinguishable from the sands of the other, both being probably of fresh- or brackish-water deposition.

In Mexico the yellow clays and sands of the Oligocene are also overlain by similar yellow clays and sands of Miocene age and the deposits themselves are of very similar character and thus far no evidence of a decided break in sedimentation between them has been found, the only change being in the fossils contained in them, those of the Miocene being characteristically developed in the beds around Tuxpam.

TAMPICO EMBAYMENT

The importance of the Tamaulipas Range as a structural feature is, therefore, clearly shown, and it is now apparent that at the beginning of the Tertiary and through the Eocene the waters of the Mexican Gulf covered only that portion of its present coast line in Mexico which lies north of Tordo Bay about 225 miles south of the mouth of the Rio Grande.

It would also appear that the waters of the Gulf and those of the embayment south of the Tamaulipas Range, although they were evidently separated by a comparatively narrow strip of land, did not have such a connection as would permit a commingling of the two faunas, since they have only four or five forms in common.

The Topila well mentioned above is less than twenty miles from the present shore line of the Gulf of Mexico and is much nearer to its present waters than is any bed of similar age in the region north of the Tamaulipas Range. There is no indication of a barrier to prevent the extension of these deeply buried Alazan beds with their Pacific fauna eastward or southeastward until they actually underlie the present waters of the Gulf. It is clearly evident, therefore, that during the Middle Eocene the Gulf waters did not reach below Tordo Bay on its present coast line, but were

held by a barrier or the southeastward extension of the Tamaulipas Range and that part of its present coast in the vicinity of Tampico was occupied by the waters of the Pacific.

The observed facts indicate that during the Middle Eocene the waters of the Atlantic occupied the territory north of the Tamaulipas Range and the waters of the Pacific the region south of that range, and that with the beginning of the Oligocene, the waters of the Atlantic claimed both of these regions. It is therefore evident that the connection between the Pacific Ocean and its Tampico embayment was closed entirely, either during the Upper Eocene or at its close, and that the extension of the present Gulf coast southward from Tordo Bay dates from this period.

While the exact area embraced in this embayment south of the Tamaulipas Range is unknown, its point of connection with the Pacific Ocean was probably near the present Isthmus of Tehuantepec and its western border was formed by the escarpment of the Cordilleras. It is probable that its northerly extension occupied the site of the valley now traversed by the railroad between Tampico and Monterey. The Tamaulipas Range must have formed its eastern boundary in this northern portion, but we have, at present, no idea of its eastern limit farther south beneath the western margin of the present Gulf of Mexico. It seems probable, however, that this Tampico embayment was a body of water not unlike the Gulf of California, both in extent and in trend, and that it made a peninsula of that part of the Republic of Mexico north of the Isthmus of Tehuantepec just as the Gulf of California now makes the peninsula of Lower California.

THE STRENGTH OF THE EARTH'S CRUST—*Concluded*

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PART VIII. PHYSICAL CONDITIONS CONTROLLING THE NATURE OF LITHOSPHERE AND ASTHENOSPHERE

SECTION B

RELATIONS WITH OTHER FIELDS OF GEOPHYSICS

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RELATIONS WITH OTHER FIELDS OF GEOPHYSICS

Erroneous Conclusions Reached by the Rectilinear Projection of Surface Conditions

Geologists early became aware that temperature increased with depth. Projecting this gradient as a straight line indicated that at no great depth the temperature was sufficiently high to melt all rocks and, in testimony, volcanoes brought such melted rocks to the surface. The earth was consequently looked upon as a molten or even gaseous body enveloped by a thin crust of solid rock. The logic of this conclusion seemed incontrovertible and moreover it was in accord with the simpler expectations from the nebular hypothesis. Nevertheless, direct and positive evidence from several independent sources has forced on geologists the belief that the earth is not only solid throughout, but, as a whole, is more rigid than steel. Slowly and with difficulty the older view has therefore had to be abandoned. Yet it continually recurs in one form or another, advocated chiefly by writers who see the direction in which the surface evidence of temperature gradient leads, who regard it as compulsory, and who do not recognize or give equal

weight to the direct evidence regarding the nature of the earth's interior. Because of the ease and certainty of laboratory studies there is a tendency to treat the interior of the earth as though it were incapable of speaking for itself through the evidence of geophysics, geodesy, and geology, but must remain forever a playground for the speculative imagination. Largely unknown the nature of the earth's interior is and long must be; laboratory studies on the influences of heat, of pressure, and of chemical composition, upon the physical state of the crust, must constitute the paths which guide in the search downward into the unknown; but the final test of hypothesis must be the direct testimony of the earth itself.

The rectilinear projection of surface conditions is based on the assumption that the temperature gradient is a straight line to great depths, or that strength, or density, or porosity, as the case may be, is not changed by the pressures of the interior. Such assumptions lead to views more or less in opposition to those reached in the present investigation. They must, therefore, be discussed to some degree. An illustration of these dangers of reasoning by unchecked extrapolation is supplied by a paper written by Arrhenius,[1] selected for discussion because of the eminence of the author in the fields of physics and chemistry, the definiteness with which his conclusions are stated, and the wide citation which this paper has achieved.[2] In this paper the arguments are given in favor of a gaseous nature of the interior of the earth, carrying forward an idea first suggested by A. Ritter in a series of "Researches on the Height of the Atmosphere and the Constitution of Gaseous Heavenly Bodies."

From the rate of increase of temperature with depth Arrhenius argues that at a depth of 40 km. the crust must pass into a molten condition, but one which, because of pressure, is a viscous and highly incompressible liquid. At a depth of some 300 km. the temperature, he states, must be above the critical temperature of

[1] "Zur Physik des Vulkanismus," *Geol. Foren. i Stockholm, Forhandl.*, XXII (1900), 395–419.

[2] See for example its presentation by A. Geikie, *Text Book of Geology*, Vol. I (1903), pp. 71–74.

all known substances, and therefore the liquid magma passes into a gaseous magma extending to the center of the earth. The author then notes that the chemical elements of highest atomic weight are not detected in the sun, but states that without doubt they occur, and concludes from this that they must be concentrated by virtue of gravity toward the sun's center.[1] The high density of the earth's interior is accordingly to be explained by the presence of substances heavier than surface rocks. For many reasons, as the dominance of iron in nature, as shown by meteorites, by the spectrum of the sun, and by the magnetism of the earth, it is to be concluded that this substance which he thinks necessary to account for the high density of the earth's interior is metallic iron. The earth consists consequently of the following portions measured from the center on the radius. Eighty per cent of the radius is gaseous iron, 15 per cent is gaseous rock magma, about 4 per cent is fluid rock magma, and somewhat less than one per cent is solid crust.[2]

To reconcile these conclusions with the incontrovertible evidence of rigidity, Arrhenius takes up another line of rectilinear extrapolation and carries it to an equally extreme degree. Fluids in general show a somewhat readier compressibility than solids. At high pressures then it is argued that liquids will customarily occupy less volume than solids and the pressure will tend to lower, not raise, the melting-point. Consequently, the rigidity cannot be accounted for by the maintenance of solidity through pressure. The author then points out that under enormous pressures all substances, even gases, must become highly incompressible; and that at high temperatures, where the volume is maintained the same, the viscosity of gases or fluids increases with increase in temperature. From this it is argued that in the central parts of the earth gaseous iron is more incompressible and viscous than solid steel. It is by enormous pressure consequently in spite of a gaseous nature that the interior of the earth exhibits its great rigidity.

Vulcanism according to Arrhenius is connected with the free seepage of ocean water downward through the crust which, he holds, constitutes a semipermeable membrane. By the absorp-

[1] *Op. cit.*, p. 402. [2] *Op. cit.*, p. 405.

tion of water into the heated rocks the conditions for volcanic activity are initiated. This argument, like the others, is in the form of a great extension or extrapolation of factors operative in a small way in the laboratory to conditions in nature which are wholly different in magnitude.

As comments upon this paper, it should be noted that nearly every conclusion applying to the sun and earth may be questioned.

At a depth of 1,000 km., according to Arrhenius, the temperature is about 30,000° C. The gradient is thus taken as essentially a straight line from the surface downward. There is no demonstration as to why this rectilinear extension is assumed, whether it is to be regarded as an adiabatic temperature curve produced by condensation under pressure or produced in some other way. The influence of cooling through geologic time in changing the outer gradient is not considered; nor the influence of rising magmas. The existence of radioactivity was then just beginning to be appreciated and naturally could not have been evaluated, but the data for a discussion of the other factors, though at hand, was neglected.

There is no demonstration that the heavy elements are concentrated in the sun's interior, or that the earth is mostly metallic iron. It is possible that the earth is thus constituted, but it must be proved on better evidence than a citation of the dominance of iron in nature. The incompressibility of all substances, both fluid and liquid, increases greatly with great increase of pressure following apparently parabolic curves. Therefore, it cannot be argued with any assurance that the high incompressibility of the earth's interior proves the presence of iron, or that under such pressures the fluid occupies less volume than the solid state.

At a depth of 1,000 km. Arrhenius states that the temperature is about 30,000° C. and the pressure 250,000 atmospheres.[1] If, under these conditions of exalted temperatures, gaseous rock or iron has a viscosity equal to that of solid steel it may well be asked how the stars, with their immensely greater masses and consequent internal pressures, can maintain a convective circulation competent to keep up their enormous surface radiation. Furthermore, however viscous a compressed gas or liquid may be, this property

[1] *Op. cit.*, p. 400.

should be distinguished from rigidity. If a body can resist even small shearing stresses for an indefinite period, it has the essential properties of a solid and not a gas. If it possesses real rigidity, even if it should be true that under relief from pressure the substance would turn into a gas, yet such relief cannot take place and it is a confusion of terms to speak of the substance as a gas when exhibiting to a striking degree the essential qualities of a solid. This distinction between viscosity and rigidity is of first importance, yet is not mentioned by Arrhenius. Although undercooling of a fluid into a glass gives rise to the elastic properties of a solid, it has not been shown that increase of pressure, however great, upon a gas above the critical temperature would transform increasing fluid viscosity into solid rigidity and plasticity such as is exhibited by the earth.

As to the hypothesis that the crust is a semipermeable membrane, permitting a free downward seepage of ocean water, but little need be said, since this is a subject which has been much discussed in recent years and is now largely discarded by geologists. The evidence against it is varied. Petrologic study shows the deep rocks to be impermeable and unaltered; beyond a shallow depth they are dry, and their gaseous and liquid occlusions are held unchanged for geologic ages. Unsound conclusions have been built upon the behavior of steam within porous sandstones, combined with confusion of the rate of diffusion under enormous pressure-gradients in the laboratory with enormous pressures, but low pressure-gradients within the crust. Furthermore volcanoes are not restricted to the vicinity of the sea and their emanations are not of the proper composition to have been derived from ocean waters. As Suess has said, volcanoes are not nourished by the sea, but every volcanic eruption adds to the waters of the ocean.

The paper under discussion was written by a scientist who has done much exact work in physical chemistry, but who in passing to geologic thinking has adopted the habit of an earlier generation —a habit of speculative thought, suggested by chemical and physical concepts and not verified by a study of the earth. The form of present geologic investigations has, however, advanced to the quantitative stage, although the data are often so inexact that the

order of magnitude, or the direction of the truth, is all which may be now ascertainable.

In conclusion, it is seen that the hypotheses outlined by Arrhenius imply a thinness of the crystalline lithosphere and a crustal weakness wholly at variance with the conclusions regarding strength which have been reached in this investigation. They imply a difference in nature of the earth's interior from that given by the more direct lines of evidence, as shown by the body resistance of the earth to vibratory distortions of both short and long periods. Because of these many difficulties, this group of hypotheses, adopted by Arrhenius, has already been largely discarded, though they still find considerable acceptance, more especially by workers in related fields of science. But the measures of lithospheric depth and strength which appear to be given by geodesy add their testimony to the cumulative evidence against these views.

THE EVIDENCE OF TIDES ON RIGIDITY AND STRENGTH

The tidal distortion of the solid earth measured by means of the horizontal pendulum has shown that its rigidity is of the order of magnitude of steel. But the recent measurements by Michelson and others, employing a long horizontal pipe partly filled with water, showed clearly that the earth's rigidity is even greater than that of steel.[1] This higher value is in agreement with the inductions from the observations on the variations of latitude. But these measurements give the rigidity of the earth as a whole, not the distribution of rigidity. The resistance to tidal deformation is furthermore complicated by the influence of gravity and increasing density with greater depth. Even if the earth were a liquid globe it would resist tidal distortion to one-third the degree of the resistance of a globe of steel, and if the liquid sphere were denser inside, this ratio would be further decreased.[2] Notwithstanding this factor, however, it is clear that the earth as a whole is more rigid than steel. As the outer part is known to be less rigid than steel, it follows that the rigidity of much of the interior must be

[1] "Preliminary Results of Measurements of the Rigidity of the Earth," *Jour. Geol.*, XXII (1914), 118.

[2] A. E. H. Love, *Elasticity*, p. 306.

proportionately higher. But the tidal stresses, though serving as a measure of the rigidity of the earth as a whole, are so small that they are ineffective as a measure of the strength of the earth as a whole, or of even its weakest parts. The smallness of the stresses can be appreciated by noting Darwin's numerical calculations. In his original paper Darwin arrived at the conclusion that the tidal stress-differences at the center of the earth were eight times as great as at the surface, and this result has been widely quoted. In the final publication, however, a correction is made showing that this is the ratio between the surface stress *at the poles* as compared to the center. The stresses at the poles, at the equator, and at the center he finds to be in the ratio of 1 to 3 to 8. The diurnal tide gives an actual stress-difference per square centimeter amounting to 16 grams at the poles, 48 at the equator, and 128 at the earth's center.[1] The strength of granite at the surface of the earth averages about 1,700,000–2,000,000 grams per square centimeter. The elastic limit for steel subjected to tensile or compressive stresses in one direction ranges from about 3,500,000 grams to 4,500,000 grams per square centimeter, according to the grade of the metal. The ultimate strength is about twice as high as the elastic limit. Thus the earth is stressed by the tidal forces even at the center to only about one part in fifteen thousand of the strength of good granite at the surface, or about one part in twenty-seven to thirty-five thousand of the limits of perfect elasticity which steel exhibits in the laboratory. With stresses so small it is not surprising that although tides give measurements of rigidity their evidence regarding viscosity is most uncertain. The results of estimates of the viscosity are more or less contradictory and so small as to be within the probable error of determination. Nevertheless Schweydar considers that there is a suggestion of a slightly plastic zone extending from a depth of about 120 to 620 km. Although this has been adopted in the present article as the limit of the asthenosphere, it would appear that the convincing proof for the existence of such a zone, and the determination of its limits

[1] George H. Darwin, "On the Stresses Caused in the Interior of the Earth by the Weight of Continents and Mountains," *Collected Scientific Papers* (1908), II, p. 481; original publications, *Phil. Trans. Roy. Soc.*, CLXXIII (1882), 187–223, and *Proc. Roy. Soc.*, XXXVIII (1885), 322–28.

also is more likely to be given by the geologic and geodetic evidence rather than from that yielded by the tides, provided that the present hypothesis of the existence of an asthenosphere is accepted.

It might seem that if the asthenosphere is strained to its limit by permanent stress and is slowly yielding, that even the small and rhythmic tidal stresses, like the last straw on the camel's back, might reveal a lack of resilience in the region of yielding. The distinction was emphasized in Section A, however, that an elastic limit which is determined for permanent stress by a facility of recrystallization at a high temperature may be a far lower elastic limit than that which would exist for rapid rhythmic stresses. Recrystallization would theoretically go forward a little more rapidly during the additive phase of the tidal stress, but the process is presumably so slow, and the tidal stress so small and rapid, that no appreciable effects would be attained before the following of the negative phase. A high resilience of the earth under tidal stress seems therefore quite compatible with the existence of a slowly yielding asthenosphere.

THE EVIDENCE OF EARTHQUAKE WAVES ON RIGIDITY AND DENSITY

The speed of an elastic wave through a solid varies directly with the square root of the modulus of elasticity and inversely with the square root of the density. There are two waves, corresponding to the elasticities of volume and form respectively, the one measured by the modulus of compressibility, the other by the modulus of rigidity. The first is the longitudinal or radial wave, the second is the transverse wave. The former outruns the latter and gives rise to the first preliminary tremor by which the earthquake records itself in distant regions. The transverse vibration is felt as the second preliminary tremor, followed by the much larger oscillations of the principal wave. The first two go through the earth, the latter passes around the surface. The fact that there is a transverse wave shows that the earth is solid throughout. But the vibrations at the point of emergence for waves which have penetrated more than half-way into the earth are so faint because of distance that their beginnings are in doubt, and consequently the speeds of transmission below one-half of the radius are uncer-

tain. These greater depths do not, however, so immediately concern the present subject. For the outer quarter of the earth both radial and transverse waves increase in velocity of transmission with depth, showing that incompressibility and rigidity increase faster than density and reach values greater than those exhibited by steel at the surface of the earth.[1]

So much is certain, but when it comes to testing the character of any particular shell by means of the velocities and character of the vibrations which have passed through it, there is but little certainty. The difficulty of an exact interpretation is discussed well by Knott.[2] To illustrate the variety of opinions, Benndorf has worked out a law according to which the speed of transmission increases rapidly to a depth of 200 miles (320 km.) from the surface. Knott assumes a constancy of speed below a depth of 400 miles (644 km.).[3] Wiechert has concluded that there are sudden changes in velocity at depths of 1200, 1650, and 2450 km. Poisson's ratio which expresses the relationships of the elasticities of form and volume remains, however, practically constant throughout, having a mean value of 0.27.[4] These changes imply surfaces of discontinuity. If real, however, they are deeper than the shells of the earth involved in the problems of isostasy. The conclusions rest, however, upon data of doubtful reliability. Reid has made a critical examination of this subject in connection with his comprehensive study of the excellent records obtained from many parts of the world of the California earthquake of 1906.[5] Following Wiechert's method, the curves representing the normals to the wave fronts and the velocities at various depths were computed from the data of the seismograms. The result showed that for the radial or longitudinal wave the velocity increased rapidly with depth but with decreasing rapidity, from 7.2 km. per second at the

[1] Galitzen, *Vorlesungen über Seismometrie*, p. 138, 1914.

[2] *Physics of Earthquake Phenomena* (1908), chap. xii.

[3] *Op. cit.*, pp. 248–50.

[4] G. W. Walker, *Modern Seismology*, 1913, p. 61.

[5] *California Earthquake of April 18, 1906* (Report of the State Earthquake Investigation Commission, Vol. II, "The Mechanics of the Earthquake," by H. F. Reid). Published by the Carnegie Institution of Washington, 1910.

surface to 12.5 km. per second at 2,170 km. from the surface, 0.66 of the radius from the center. Below that depth the velocity is nearly constant. The velocity of the transverse waves is 4.8 km. per second at the surface and increases almost linearly with depth, reaching a velocity of about 7.5 km. per second at half the distance to the center of the earth. The absence of good records from distances beyond 125° prevents a knowledge of the velocities at greater depths. Within the limits regarding which information is given, Reid remarks that there is no indication of a sudden change in the velocity of either wave such as we should expect if there were any sudden changes in the nature of the earth's interior. Oldham also finds no evidence of sudden change to a depth of at least 2,400 miles, 0.4 radius from the center.[1] From the curves showing the relation of velocity to depth which Reid gives[2] it is seen that the ratio of velocity of the transverse to the velocity of the longitudinal wave is 0.66 at the surface, 0.56 at 0.95 R, 0.53 at 0.9 R, reaching a minimum of 0.52 at 0.85 R, from which it increases to 0.58 at 0.5 R. This shows that both moduli of elasticity increase with depth, but that down to a depth of between 0.8 and 0.9 R. from the center of the earth, 637 and 1,274 km. from the surface, incompressibility increases relatively faster than rigidity. The change is shown as very rapid in the first 300 km. This is the only way in which the existence of an asthenosphere reflects itself in the rigidity of the earth, and this may not be related to its weakness but to some other property, such as the nature of compressibility or of changing chemical composition, or partly in the lack of detailed knowledge in the nature of the data.

Earthquake waves, like the tides, measure elasticity rather than strength. The vibrations which penetrate 200–300 km., and more, downward in the earth are already greatly reduced in amplitude and therefore in the strains which they bring on the earth. What the maximum strains may be is unknown, but reasonable assumptions as to amplitude show that within the asthenosphere the order of magnitude of the strains would be of the nature

[1] "On the Constitution of the Interior of the Earth as Revealed by Earthquakes," *Quar. Jour. Geol. Soc.*, LXII (1906), p. 470.

[2] P. 122.

of a thousandth part of that which granite at the surface of the earth can sustain. Furthermore, even if the stresses were greater and could be used as a measure of strength, this would apply to sudden stresses only and the results obtained from elastic vibrations could not be used safely as a means of determining the strength under long-enduring stresses. Thus the evidence from both tides and earthquakes is negative in regard to the existence of an asthenosphere. They show only that it is not fluid and that it is not markedly unlike the rest of the earth in its elastic properties.

HIGH, BUT VARIABLE, ELASTIC LIMIT WITHIN THE UPPER LITHOSPHERE

The experiments by F. D. Adams showed that under conditions of cubic compression rocks became far stronger than when subjected to compression, as at the surface of the earth, in one direction only. When a cylinder of Westerly granite was incased in a steel jacket and then subjected to heavy pressure upon its ends, a small cavity within the specimen just began to break down under a stress-difference of between 160,000 and 200,000 pounds per square inch, about six to eight times the strength possessed by this rock under surface conditions. At a temperature of 550° C., a temperature calculated to exist at a depth of 11 miles below the earth's surface, small cavities remained open when submitted to considerably greater pressures than occur from the overlying load at this depth.[1]

Adams' experiments and King's calculations are most important and show without doubt that the more superficial parts of the earth, to a depth of ten to fifteen miles at least, are far stronger than had been supposed; but they apply to the temperature and pressure gradients in places of geologic quiet, not to regions undergoing igneous intrusion and crustal deformation. Then the temperatures may become far higher and the crust surcharged with magmatic gases. Yet it is under these conditions especially, of geologic activity as contrasted to geologic quiet, that regional metamorphism and rock flowage proceeds. Still less does this experimental work prove a great strength of the crust at depths of more

[1] Louis Vessot King, "On the Limiting Strength of Rocks under Conditions of Stress Existing in the Earth's Interior," *Jour. Geol.*, XX (1912), 136, 137.

than a hundred kilometers, for there the temperatures are presumably above those which under the conditions of freedom from pressure at the surface of the earth produce dry fusion. Occluded gases, furthermore, are held beyond possibility of escape.

The strength of the crust is dependent consequently upon fourfold conditions—the nature of the material, the cubic compression, the relation of temperature to the point of fusion, and the rapidity of the application of the stress. These factors are all variable with time and place. How variable will be seen upon further consideration in the following paragraphs.

The influence of the nature of the material is seen when it is noted that granite is only about one-half as rigid as the basic rocks, although it is not less strong. Consequently, regional stress coming upon a complex of two such rocks will elastically deform the granite more readily, a greater stress will be thrown upon the basic rocks, and since their elastic limit is not correspondingly higher they should begin to yield by flow or fracture before the more pliant rocks had reached their limit. The general conclusion is that a movement of compression in the earth's crust must necessarily give rise to unequal strains and concentration of stress, as well from variations in chemical composition as from variations in structure. The local stress may rise far higher than the general regional stress.

As to the second factor, during the progress of normal faulting the horizontal compressive stress in the crust is less than the vertical stress due to weight. During the progress of folding and mashing, on the contrary, the horizontal stresses become far higher. But the least of the three principal stresses determines the amount of cubic compression; the difference between the greatest and least stresses determines, on the contrary, the amount and direction of the strain upon the rigidity of the rock. Thus it is seen that both the cubic compression and the stress-difference vary with the amount and kind of forces.

It is temperature, however, which is probably the most variable of these factors. Igneous activity brings the temperatures of the greater depths comparatively near to the surface and must produce

widespread weakening of the crust, both through the physico-chemical effects of the exalted temperatures and the structural effects of the intruded viscous fluids.

The rapidity of the application of stress is a variable in itself and furthermore has variable effects, but would seem, however, to be the least important of these several factors. The movements of horizontal compression and vertical warping are slow and give time for recrystallization in the deeper crust. In this way they meet a lesser resistance than would rapid stresses. Where the temperatures are close to those of fusion it would seem in fact that rock flowage by recrystallization, developing the gneissoid structure, should demand markedly less shearing stress than the process of granulation. The gnarled and twisted rocks of the Archean speak of the presence beneath them of molten magmas rather than of an enormous degree of compressive forces upon them. But ready yielding by recrystallization in one place would permit the concentration of mashing stresses upon other localities and raise the strain to that intensity needed for granulation. An enormous depth of cover, such as Adams' experiments have been thought to show, is not suggested by the geologic evidence, nor apparently is it demanded by a completer theory.

In fault movements and in dike or sheet intrusion accompanied by the expansion of gases are two sources of rapid application of forces. It is probable, however, that their deformative action is confined to the outer ten miles of the crust, and their consideration need not detain us in the evaluation of those factors of strength which concern the crust as a whole.

Summing up the conclusions from these various lines of evidence, physical and geological, it is seen that they suggest a rapid increase of strength with depth, then the gradual passage into a deep zone of lowered strength. The limits and values, however, are variable with time and place. Such a distribution of strength as is indicated by these independent lines is in accord with the interpretation of the geodetic evidence showing the existence of crustal competence to support heavy loads over certain limits of area, coexisting with flotational equilibrium over much broader regions.

MODES OF LITHOSPHERIC YIELDING AND THEIR RELATION TO STRENGTH

The relationship of strength to depth which has been derived in this study and which was expressed in the curve of strength at the end of Part VII is to be connected with the physical qualities discussed in this part. Here it is seen that it is a curve of elastic limit. When that limit is exceeded, permanent deformation must take place; by one means at the surface, by another within the body of the lithosphere, by still another at its base.

At the surface the typical mode of yielding is by jointing and faulting, in stratified beds by folding also. The movements in this zone of fracture and in the transitional zone of combined fracture and flow may be regarded as merely the responses in a thin, brittle, and relatively weak outer layer to deformative movements progressing in the great thickness of the lithosphere below. But the rocks of deeper origin which have been exposed at the surface by profound erosion show that they have yielded in another fashion. Their foliated structures and crystalline textures testify to yielding by massive flowage. Fracturing appears to have been absent, except in so far as it was produced by intrusions from below, giving rise to complexes of dikes and sheets. These visible exposures suggest that at still greater depths, notwithstanding the great strength of that zone, open fracture planes disappear and rock flowage both by granulation and by recrystallization is still more distinctive. This appears then to be the mode of yielding of the great body of the lithosphere.

Recently Becker has suggested that fracturing may enter into the problem of isostasy in the following way: The demonstrated capacity of small cavities to remain open under great pressures may permit fissuring and jointing to extend deeper into the crust than had been previously thought possible. To the degree to which fractures and porosities do exist they must decrease the specific gravity of rocks. If shattering pervaded the rocks of one region and not another, even though the rocks were exactly alike in composition, the densities would become different. To give isostatic equilibrium the region of shattered rocks would have to stand higher than the other. This would be the initial effect as

a result of the decrease in density, even if the zone of compensation rested on an unyielding base.[1] The logical correctness of this argument is not to be questioned, but rather the degree of its application. The following arguments suggest that shattering or porosity are, however, very subordinate rather than determining factors in the isostatic problem.

Such a theory does not account readily for the movements needed to maintain isostasy because of erosion and sedimentation. These surface changes of mass suggest a restoration of mass by lateral undertow. Furthermore, the appeal to nature shows that the rocks, once deep-seated, which have become revealed at the surface by erosion, are almost without pore space. The average porosity according to Fuller is 0.2 per cent, but the mean differences in densities between ocean and continent which must be accounted for under the hypothesis of uniform compensation to a depth of 122 km. amount to about 4 per cent.

Joints are observed to decrease with depth, becoming tighter and more distantly spaced, and the indications given by the lack of general circulation of ground-water through crystalline rocks, except within joint spaces near the surface, are that at greater depth the joint spaces are negligible.

In the great compressive movements the whole thickness of the crust must yield, but even this cannot be conceived as producing porosity by granulation sufficient to notably modify the density. A large part of the deformation in the deeper crust must be by a process of recrystallization. Assume, however, that granulation is the dominant process. Observation of granulated rocks shows a reduction in size of the crystals, but these broken fragments fit against each other perfectly and without great internal distortion of crystals. In granulated rocks from the zone of flow there is therefore always some amount of recrystallization, sufficient to eliminate that porosity connected with minute shattering and movement of the broken particles. The explanation appears to be as follows: The minute shattering of the minerals tends to give a high pore space, but with a high pore space the amount of contact

[1] G. T. Becker, "Isostasy and Radioactivity," *Science*, XLI (1915), 157–60; "On the Earth Considered as a Heat Engine," *Proc. Nat. Acad. Sci.*, I (1915), 81–86.

between grains becomes proportionately less. For the prevention of ready recrystallization and the maintenance of this pore space the granulated rock, according to present theory, must be conceived of as dry and the grains accordingly unsupported except at the points of contact. The shear strains within each grain become very great in proportion to the diminution of contact, and increase in proportion to the regional pressure. If the points of contact, for example, cover only one-fourth of the surface, the compression on those points would be four times as great per unit of surface as if there were continuous contact between grains. On the intervening parts of the surface there would be no pressure Internal shears would result in this way from the hydrostatic pressure of dry rock due to depth and are not dependent upon a pressure-difference in the rock as a whole. The internal strains would tend to produce molecular changes of state as in the plastic flow of metals. There would be melting to relieve the strain, and refreezing by which the molecules would build out the crystals into the pore spaces. By this means recrystallization can go on without the aid of crystallizers, though presumably with more difficulty, and the comminuted crystals come to fit compactly as they are observed to do. This elimination of porosity presumably goes on approximately with the process of granulation, though it may lag somewhat. It would go forward more effectively with depth, irrespective of temperature, since there would be the greater static load upon the rock and the greater differential pressures within the mineral particles. It might be expected that such reduction of pore space would go forward to a limited extent only, leaving a residual porosity. Observation, however, shows that the pore space has been almost completely eliminated. Furthermore, the rocks now exposed at the surface acquired their absence of pore space at depths of only a few miles from the surface. At depths measured in tens of miles there seems then no expectation that density would be notably decreased because of a development of porosity.

To sum up the modes of yielding within the lithosphere: at the surface is seen to exist a thin outer crust intimately cracked on the outside by closely spaced parallel joint systems. Local

extreme deformation is by faults and folds. With increasing depth and strength the joints become less abundant and faults pass into flexures. The passage of fractures into flexures implies the beginnings of massive flow. Where magmatic heat or emanations are not present the mode of mashing is presumably more especially by granulation. With still greater depth the yielding becomes more uniformly distributed throughout the rock mass. Both because of this pervasiveness of mashing and the great strength of this zone, deformation here requires the most force and absorbs the most energy of any part of the lithosphere. At greater depths the rock is more compressed, and is still more rigid than above, but the temperature here approaches fusion; recrystallization readily takes place, the strain which can be elastically carried is in consequence low, and the lithosphere passes gradually into the asthenosphere. Where, however, magmas rise through the crust they carry with them the environment of the asthenosphere; the lithosphere becomes locally abnormally heated and saturated with magmatic emanations. Recrystallization goes forward readily and the zone of weakness penetrates upward even to the zone of fracture. Thus in the injected and crystallized roofs of ancient batholiths, laid bare by profound erosion, we may perceive the nearest approach to dynamic conditions which prevail in depths forever hidden.

POPOSAURUS GRACILIS, A NEW REPTILE FROM THE TRIASSIC OF WYOMING

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Some time ago the writer mentioned a new form from the Triassic of Wyoming and briefly described one of the vertebra.[1] At that time it was hoped that the explorations in the western Trias of the past season would bring to light something further concerning this peculiar form. No additional evidence has so far been disclosed, however, in the collections from Arizona and New Mexico, and it is probably well at this time to describe more fully the remains for the benefit of others who may be engaged in the same field.

The material herein described consists of an ilium, several vertebrae from various parts of the column, two femora, a tibia, the proximal end of a fibula, and several innominate fragments. All are considerably crushed and broken. The leg bones are so flattened that it is rather difficult to determine the true proportions.

The ilium, as previously pointed out,[2] is very similar to that described by J. H. Lees[3] as belonging to *Paleorhinus Bransoni*. The condition of the bone suggests little distortion; still, upon a comparison with the specimen described by Lees, it is evident that the supra-acetabular flange has been somewhat depressed by pressure. To all appearances the acetabulum is formed almost entirely by the ilium. It is broad and deep and closely confined above by the supra-acetabular flange. In preparing the specimen what appears to be a foramen of perhaps 5 mm. in diameter was exposed, running in and slightly upward from the upper acetabular surface. It is barely possible, however, that this is simply a matrix-

[1] M. G. Mehl, "The Phytosauria of the Rocky Mountain Trias," *Jour. Geol.*, 1914 (in press).

[2] *Op. cit.*, p. 159.

[3] "The Skull of *Paleorhinus*, a Wyoming Phytosaur," *Jour. Geol.*, XV (1905), 44.

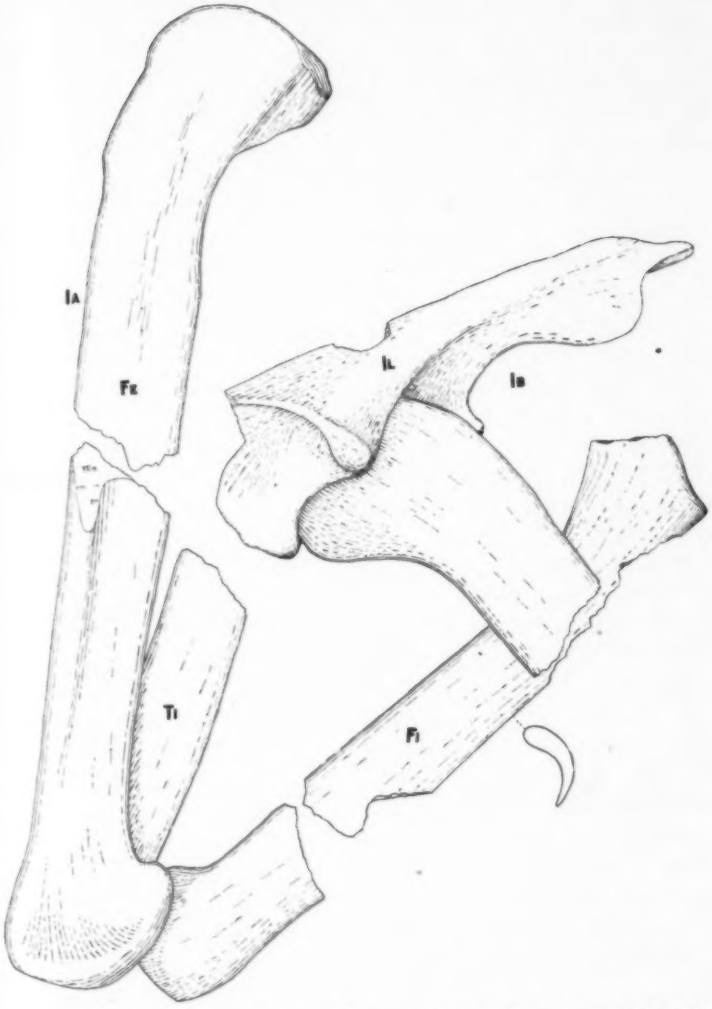

FIG. 1.—*Poposaurus gracilis.* 1*a*, left femur, tibia, and fibula as they lay in the matrix. 1*b*, right ilium and femur. All about ½ natural size.

filled crack. From the center of the margin of the supra-acetabular flange a strong ridge runs up and back in a regular curve. Below, this ridge is about 20 mm. in width from which it gradually narrows till it becomes confluent with the upper posterior margin of the ilium. The upper posterior process is broken away, but the indications are that it was a thin, platelike process with rounded border, separated from the lower posterior process by a deep sinus. The upper anterior process is irregular in shape, extending forward in a regularly expanding spatulate form. A marked ridge extends along its outer surface from the supra-acetabular flange, becoming fainter anteriorly and finally confluent with the upper anterior border. The process makes up nearly half of the total length of the ilium, the greatest length of which is 250 mm. The lower anterior projection of the ilium has a somewhat antero-posterior elongated, convex face for the articulation of the pubis. The articular face for the ischium, the lower posterior process, is somewhat larger, triangular in outline and nearly flat. These two articular faces are separated by the thin, apparently complete lower border of the acetabulum, strongly suggesting a perforate acetabulum, as in the dinosaurs.

Of the dorsal vertebrae but two are sufficiently well preserved to be of use in description. They probably represent the anterior thoracics inasmuch as they show an early stage in the transition of capitular facet from the anterior face of the centrum to the neural arch. The description of one of these will suffice as they show essentially the same features.

The centrum is spool-shaped, moderately biconcave, and very much constricted laterally between the articular faces. This has probably been greatly accentuated by pressure. The articular faces are oval in outline, about 32 mm. wide and 39 mm. high. The centrum is 54 mm. long and was probably 15 mm. wide at the center in an uncrushed condition. The neural arch is high and delicately constructed. The diapophysis is a thin horizontal plate confluent anteriorly with the articular face of the zygapophysis and gradually expanding posteriorly. At its posterior extremity it suddenly thickens below for the tubercular facet which is separated from the posterior zygapophysis by a deep, rounded sinus.

At this point, the greatest width of the diapophyses, the two tubercular facets are approximately 50 mm. apart. Near the mid-length the diapophysis is supported below by two thin diverging buttresses, the anterior one confluent below with the upper anterior

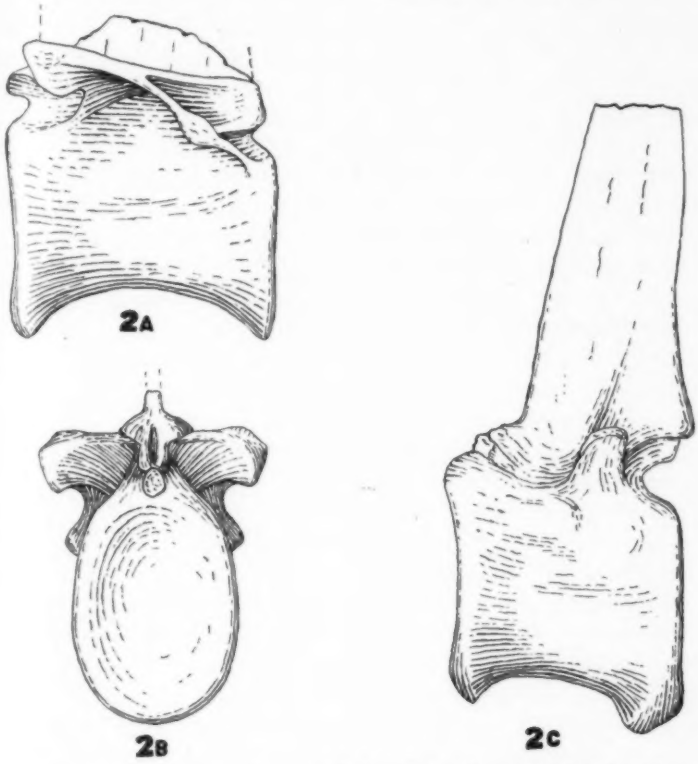

FIG. 2.—*Poposaurus gracilis*. 2*a*, right lateral view of one of the dorsal vertebrae. 2*b*, posterior view of the same. 2*c*, left lateral view of second caudal vertebra. All about ⅔ natural size.

face of the centrum. Just before reaching this point it is considerably swollen to form the capitular facet for the reception of the rib. The posterior support is directed toward the upper posterior face of the centrum, but loses its identity before reaching that

point. The distance between the capitular and tubercular facets is about 38 mm. The articular faces of the zygapophyses are of moderate size. The anterior pair are directed up and in; the posterior pair, of necessity, out and down. The anterior pair form a cuplike depression with a narrow antero-posteriorly directed median sinus into which apparently extended a thin vertical support of the posterior pair of zygapophyses of the preceding vertebrae, a condition that would apparently restrict somewhat the free motion of the vertebral column. But little remains of the spine. At the base it is 5 mm. thick and about 45 mm. wide.

The sacrum consists of at least four, and probably five closely united vertebrae with a total length of perhaps 230 to 290 mm. It is considerably abraded and lacks most of the neural arch. The centra of the component vertebrae rapidly decrease in size posteriorly. The anterior face of the first is about 45 mm. in diameter. All the centra are considerably constricted laterally. Apparently only the anterior two vertebrae bear sacral ribs, though this is not certain. The ribs were stout, somewhat flattened, and articulated low on the neural arch near the mid-length of the vertebra.

The caudal vertebrae are represented by the first and second of the series. The second is nearly perfect and will serve well to characterize the group. The centrum is 50 mm. long, somewhat more deeply biconcave than the dorsals, and much constricted laterally at mid-length. The articular faces are some 40 mm. wide and 50 mm. high. Below there is a strong inter-centrum articulation for the reception of chevron bones. These articulations are well developed on both the anterior and posterior faces, but considerably stronger on the anterior face of the vertebra. The diapophyses are placed high on the neural arch and well back. Their tips are slightly higher than the neural canal. They are short, directed upward, and slightly back. From tip to tip they measure 42 mm. The zygapophyses are placed close together, the anterior ones directed in and upward. The spine is long and thin and directed very slightly backward. At the base it is about 5 mm. thick and 35 mm. wide. The extremity is broken away, probably no considerable portion, leaving the total height of the vertebra 135 mm.

The first caudal differs from the second in that the centrum of the former is somewhat longer, has larger articular faces, and has the chevron articulation at the posterior end only. There is also a considerable difference in the diapophyses of the two vertebrae. In the first they are broad and, although badly weathered, suggest a lateral face of perhaps 30 mm. anterior-posterior extent as compared with 10 mm. in the second. In the first caudal the diapophyses are placed farther forward and slightly lower than in the second and are slightly supported by the centrum.

The condition of the leg bones is such that little can be done in the way of description other than stating approximate measurements. The femur is at least 465 mm. long and very slender. In an uncrushed condition the shaft was probably about 45 mm. in diameter. The head is considerably expanded antero-posteriorly, measuring 85 mm. in that direction. Little can be said of the condyles except that the antero-posterior extent of the articular face was great, perhaps 85 mm. or more.

A piece of the tibia, 125 mm. in length, a portion of the shaft near the proximal end, remains imbedded in the matrix in its natural relation with the left femur. In life the shaft was probably about 30 mm. in diameter. The left fibula, some 345 mm. in length, is very thin laterally and much expanded antero-posteriorly at both the proximal and distal ends. The shaft, which is 40 mm. in width, is regularly concavo-convex antero-posteriorly throughout its entire length, the convexity outward. Were this a feature due to distortion only, one would hardly expect the solid articular ends to conform to the curvature of the thin shaft as they do. Furthermore, polished cross-sections of the shaft do not suggest considerable flattening of its walls.

RELATIONSHIPS

The specimen has little in common with the phytosaurs; besides its apparently perforate acetabulum, of which there is but the slightest suggestion in some of the phytosaurs, it shows a marked difference in the shape of the ilium and in the comparatively large number of vertebrae fused together in the sacrum. In the phytosaurs there are but two sacrals and these are free. As *Dolichobrachium*, described by Dr. S. W. Williston from the Trias of

Wyoming,[1] is known from the humerus, teeth, and coraco-scapula only, there is little basis for comparison. The first two elements and especially the coraco-scapula are suggestive of a more massive build than that of the present specimen. There is, however, the possibility that these two forms are identical, an uncertainty that must remain till more abundant material is to be had. In some respects it resembles some of the early dinosaurs. The hollow, slender leg bones, the slightly biconcave vertebrae, are suggestive of the Theropoda and the perforate acetabulum is typically dinosaurian. Unlike the condition in this group of dinosaurs, however, each sacral rib is supported by a single vertebra. While this is a condition found in the sauropod dinosaurs, their solid bones, the expanded neural canal of the sacrum, and many minor points are very different from the form under discussion. In the future it may be shown that the genus *Palaeoctonus* of Cope is related to the form here described as, indeed, there is a possibility that *Palaeoctonus* and *Dolichobrachium* are identical. When the remains of the latter genus were discovered there were present portions of a skull in a powdery condition, too poorly preserved to be saved. According to Dr. Williston there was enough seen of the teeth, however, to suggest a possible relationship with *Palaeoctonus*.

Everything in the structure of the form so far studied indicates a well-muscled creature light in weight, possibly bipedal in gait occasionally, and most assuredly swift in movement. The name *Poposaurus gracilis* is suggested for the material herein described from its discovery on the Popo Agie river.

The type specimen is numbered 602 in the University of Chicago collections. It was collected by Professor E. B. Branson in the red beds near Lander, Wyoming.

[1] "Notice of Some New Reptiles from the Upper Trias of Wyoming," *Jour. Geol.*, XII (1904), 688.

THE CANNONBALL MARINE MEMBER OF THE LANCE FORMATION OF NORTH AND SOUTH DAKOTA AND ITS BEARING ON THE LANCE-LARAMIE PROBLEM[1]

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Field examination by the writers and the paleontological determinations by Drs. Stanton and Knowlton during the years 1912 and 1913 show that in a large region west of Missouri River in North and South Dakota the Lance formation consists of two distinct parts, a lower non-marine part containing a flora very similar to, if not identical with, that of the Fort Union and an upper marine member containing a fauna closely resembling, but not identical with, that of the Fox Hills sandstone. This upper part, on account of its peculiar fauna, has been mapped separately and named the Cannonball marine member of the Lance formation. Farther west non-marine beds bearing lignite and occupying a similar stratigraphic position have been named the Ludlow lignitic member of the Lance.

The area examined by the writers embraces a territory of over 5,000 square miles, extending from Mandan, North Dakota, west to Montana, a distance of about 175 miles (Fig. 1), and south beyond the boundary line of South Dakota. The marine member of the Lance has been mapped from Mandan to a point 4 miles west of Haley, North Dakota, a total distance of about 130 miles. A large part of the area examined by the writers has been described already from an economic standpoint.[2]

The Lance formation had previously been mapped in two adjoining regions, the Bismark Quadrangle[3] and the Standing Rock

[1] Published by permission of the Director of the U.S. Geological Survey.

[2] E. Russell Lloyd, "The Cannonball River Lignite Field, Morton, Adams, and Hettinger Counties, North Dakota," *U.S. Geol. Survey Bull. 541–g;* D. E. Winchester, C. J. Hares, E. M. Parks, and E. Russell Lloyd, "The Lignite Field of Northwestern South Dakota," *U.S. Geol. Survey Bull.* (in course of preparation); C. J. Hares, "Lignite in Southwest North Dakota," *U.S. Geol. Survey Bull.* (in course of preparation).

[3] A. G. Leonard, *U.S. Geol. Survey Geol. Atlas*, Bismarck Folio (No. 181), 1912.

Indian Reservation,[1] in both of which one of the writers (Lloyd) has been able to distinguish the upper marine member. Parts of these areas are included in the accompanying map (Fig. 1).

Throughout the greater part of the area studied by the writers the strata are nearly flat-lying, with only a very low general dip, usually less than 30 feet per mile, to the north or northeast. Practically the entire area is grass-covered and only in the bluffs of the larger streams and locally in high steep-sided buttes are natural rock exposures found. The relief is very slight; hence there is considerable difficulty in tracing the formation boundaries.

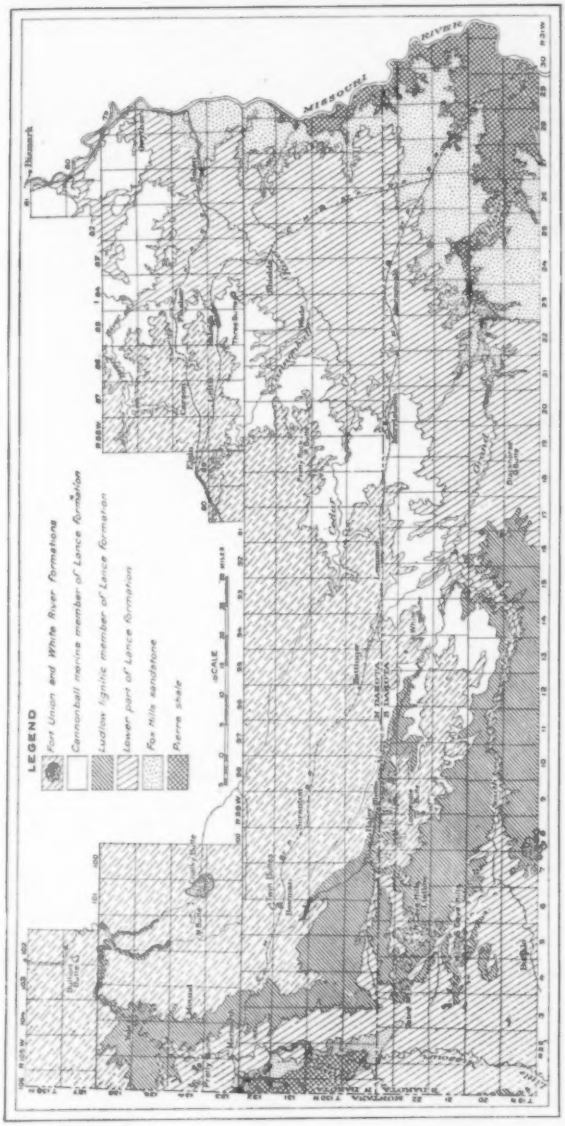
The general character and relations of the Cretaceous and Tertiary formations exposed in this part of the Dakotas are shown in Table I. The areal distribution is shown on the map (Fig. 1).

TABLE I

CRETACEOUS AND TERTIARY FORMATIONS IN WESTERN NORTH AND SOUTH DAKOTA

System	Series	Formation	Character	Thickness in Feet
Tertiary	Oligocene	White River	Cross-bedded sandstone and fresh-water limestone	140
Unconformity, angular up to 20° and erosional up to 1,500 feet				
Tertiary	Eocene	Fort Union	Yellow sandstone, shale, clay, and lignite	1,025
Tertiary?	Eocene?	Lance	Cannonball marine member. Dark sandy shale, dark shaly sandstone, and yellow sandstone, containing marine shells	0–300
			Ludlow lignitic member. Light sandy shale, calcareous sandstone and lignite	0–350
			Somber-colored shale, yellow sandstone, and thin lignite beds	400–525
Cretaceous	Upper Cretaceous	Fox Hills	Sandstone, yellow	25–400
		Pierre	Dark marine shale	300+

[1] W. R. Calvert and Others, "Geology of the Standing Rock and Cheyenne River Indian Reservations, North and South Dakota," *U.S. Geol. Survey Bull. 575*, 1914.


LEGEND
Fort Union and White River formations
Cannonball marine member of Lance formation
Ludlow lignitic member of Lance formation
Lower part of Lance formation
Fox Hills sandstone
Pierre shale
SCALE
MISSOURI RIVER
Bismarck
Mandan
Dickinson
Hebron
Glen Ullin
Hannover
Richardson
Center
Beulah
Hazen
Stanton
Golden Valley
Killdeer
Dunn Center
Manning
Halliday
Dunn
Mercer
Oliver
Morton
Sioux
Grant
Adams
Hettinger
Slope
Bowman
Billings
Stark
Hardin
Custer
Rosebud
Powder River
Carter
Fallon
Wibaux
Golden Valley
Sweetgrass
Stillwater
Yellowstone
Big Horn
Treasure
Rosebud
Custer
Fallon
Carter
Powder River
Big Horn
Yellowstone
Stillwater
Sweetgrass
Golden Valley
Wibaux
Fallon
Carter
Powder River
Rosebud
Custer
Big Horn
Yellowstone
Treasure

CRETACEOUS SYSTEM

FOX HILLS SANDSTONE

The type locality of the Fox Hills sandstone is Fox Ridge in the Cheyenne River Indian Reservation in South Dakota between Moreau and Cheyenne rivers and about 25 miles northwest of the mouth of the latter stream.[1] This region was examined in detail by V. H. Barnett in 1909, and the formation was mapped by Barnett and other geologists of the U.S. Geological Survey northward to Cannonball River.[2] The Fox Hills is made up of massive gray, yellow, and buff sandstone and banded shale, the maximum thickness in the Standing Rock and Cheyenne River Indian reservations being 400 feet and the average about 150 feet. Around the southern end of the Glendive anticline in Billings and Bowman counties, North Dakota, the Fox Hills sandstone is about 75 feet thick. It has been shown that in some localities in the Dakotas and eastern Montana[3] the overlying Lance formation is unconformable on the Fox Hills, the contact between the two being marked by erosion channels in the Fox Hills sandstone. In one locality in the Standing Rock Reservation the Fox Hills is 25 feet or less in thickness, so that either the erosion interval represented is of considerable magnitude or else the formation is peculiarly variable in thickness. This evidence of unconformity is, however, vitiated by the presence in the Standing Rock Reservation of a bed above the line of unconformity containing brackish water and marine fossils and along Little Missouri River by the gradation by alternation of beds of Fox Hills character with those of Lance character. The evidence both for and against the supposition of an unconformity of major importance at this horizon is fully discussed in the papers cited. From the stratigraphic stand-

[1] The name was first used by Meek and Hayden, *Acad. Nat. Sci. Philadelphia, Proc.*, XIII (1862) 419.

[2] W. R. Calvert and Others, *op. cit.*; T. W. Stanton, "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota, North Dakota, and Eastern Wyoming," *Am. Jour. Sci.*, 4th ser., XXX (1910) 172, 188; F. H. Knowlton, "Further Data on the Stratigraphic Position of the Lance Formation ('Ceratops Beds')"; *Jour. Geol.* XIX (1911) 358–74.

[3] F. H. Knowlton, *op. cit.*

point it appears to the writers that the local unconformities that have been observed are such as would be expected where a marine formation is succeeded by one of continental origin.

A significant feature of the Fox Hills sandstone and underlying Pierre shale in the Standing Rock and Cheyenne River Indian reservations is that the fauna found in the sandstone cannot be distinguished from that found in the upper part of the shale, so that the strata from which the Fox Hills fauna was collected include not only the Fox Hills sandstone but also about 200 feet of the underlying Pierre shale.

TERTIARY? SYSTEM

LANCE FORMATION

Lower part.—It has been previously stated that the Lance formation in this region consists of three parts, a lower group of shale and sandstone beds of continental origin and two contemporaneous upper members, one of sandstone and shale of marine origin, the other of sandstone, shale, and lignite of non-marine origin. The lower part of the formation outcrops in a wide belt of country in Morton County, North Dakota. Northward it passes below the flood plain of Missouri River a few miles below Bismarck, and to the west it has been mapped to and beyond the Montana-Dakota state line. It occupies a large area in Bowman and Billings counties in southwest North Dakota and in adjacent parts of Montana. Throughout this whole region it is essentially uniform in character, consisting predominantly of somber-colored arenaceous shale intercalated with lenticular beds of brown or buff sandstone. Beds of brown carbonaceous shale, bog iron ore, and thin lignite are conspicuous in most outcrops. All the strata of the lower part of the Lance are lenticular in character, and a section exposed at one locality is different in most of its details from one even a short distance away. Cross-bedding is common, especially in the sandstone. Irregularity of deposition is characteristic. Near Solen on Cannonball River, 9 miles above its mouth, the lower part of the Lance has a thickness of approximately 400 feet and on Little Missouri a thickness of about 525 feet.

The fossils found in the lower part of the Lance in the western Dakotas consist of plant impressions and bones of reptiles. The plant remains are not abundant, but a few good collections have been made from widely separated localities. All of the species belong to the widespread Fort Union flora. The vertebrates include turtles and several genera of dinosaurs, among which is the large-horned Triceratops, which is diagnostic of the Lance formation. The Ceratopsian remains are especially abundant near the base of the formation in the Standing Rock Indian Reservation, but on Little Missouri River several specimens of Triceratops have been found near the top of the formation and stratigraphically below the *Ostrea glabra* zone.

Ludlow lignitic member.—The Ludlow lignitic member of the Lance formation occupies a large area in Harding County, South Dakota, and has been mapped northward into Bowman and Billings counties, North Dakota, and eastward into Perkins County, South Dakota, where it merges with the Cannonball marine member. In the vicinity of Ludlow, South Dakota, its type locality, it consists of 350 feet of loosely consolidated buff and cream-colored calcareous sandstone and shale with interbedded lignite. It contains most of the lignite of South Dakota and the presence of this lignite is one of the chief criteria for considering it a distinct member of the Lance formation. Its lithologic character in South Dakota is very like, and its fossil flora so far as determined is identical with, the Fort Union. Its flora is like that of the lower part of the Lance, but its lithology is quite different. On the other hand, in North Dakota its flora has the same affinities as in South Dakota, but lithologically it resembles the lower part of the Lance, except for the presence of the numerous lignite beds. It is this variation in color and lithology of the Lance that renders its separation from the overlying Fort Union so difficult.

The following sections show the lithologic character of the Ludlow lignitic member.

COMPOSITE SECTION IN SECS. 32 AND 36, Tp. 22 N., R. 5 E., BLACK HILLS MERIDIAN, SOUTH DAKOTA

	Ft.	In.
Fort Union formation		
Sandstone, yellowish, and shale	255	0
Ludlow lignitic member of the Lance formation		
Lignite		8
Sandstone, somewhat shaly	45	0
Shale, dark	2	0
Lignite	2	4
Sandstone, light colored, grayish, argillaceous	12	0
Sandstone, buff, with ferruginous specks	15	0
Lignite		2
Sandstone, buff, fine-grained, muscovitic	11	0
Liginite		3
Sandstone	8	0
Lignite		6
Sandstone, buff	25	0
Shale, bluish	1	0
Shale, carbonaceous		10
Shale, arenaceous	5	0
Lignite		1
Sandstone, buff, soft	5	4
Sandstone	1	0
Shale, arenaceous with carbonaceous streaks	4	0
Sandstone, argillaceous	6	0
Lignite	1	0
Concealed interval	22	0
Sandstone, drab, cross-bedded, ripple-marked	3	0
Shale, arenaceous	4	0
Lignite	1	5
Shale, brown	3	0
Lignite	2	1
Shale, dark		5
Lignite		8
Shale	1	3
Lignite	1	5
Shale, brown	4	0
Lignite	2	3
	446	8

PARTIAL COMPOSITE SECTION, TP. 20 N., R. 9 E., BLACK HILLS MERIDIAN, SOUTH DAKOTA

	Ft.	In.
Ludlow lignitic number of the Lance formation		
Sandstone, yellow, medium fine-grained, capping hills	20	0
Shale, brown	3	0
Lignite	1	0
Shale, brown		6
Sandstone, yellow	16	0
Shale, brown	1	10
Lignite, dirty		8
Shale, brown	3	0
Shale, black		8
Lignite, dirty		6
Shale, brown	1	6
Sandstone, buff to yellow	12	0
Shale, brown	2	0
Shale, black		4
Lignite, fair		8
Sandstone and shale, with lignite streaks	13	6
Lignite, fair		8
Shale		3
Lignite, good	3	6
Shale, brown, arenaceous	1	0
Sandstone		
	82	7

COMPOSITE SECTION IN TP. 21 N., R. 7 E., BLACK HILLS MERIDIAN, SOUTH DAKOTA

	Ft.	In.
Fort Union formation		
Sandstone and shale	138	0
Ludlow lignitic member of the Lance formation		
Shale, chocolate-colored	2	0
Sandstone, buff, yellow, fine-grained, alternating with buff shale	9	0
Shale, chocolate-colored	3	4
Lignite	1	0
Shale, buff and yellow sandstone, alternating	16	0
Shale, chocolate-colored	3	0
Sandstone, brown, fine-grained	2	6
Shale	10	0
Lignite, Gianniatti bed	11	8
Sandstone, light, argillaceous	16	4

	Ft.	In.
Shale, brown	3	0
Lignite	4	2
Shale	13	0
Lignite	3	8
Shale and sandstone	15	0
Lignite, poor	3	4
Shale, carbonaceous		6
Shale		
	255	6

SECTION ON WEST SIDE OF LITTLE MISSOURI RIVER, SEC. 10, TP. 135 N., R. 105 W., NORTH DAKOTA[1]

	Ft.	In
Fort Union Formation		
Sandstone, light	15	2
Ludlow lignitic member of Lance formation		
Shale, dark	24	0
Limestone, arenaceous	2	10
Shale, blue	7	2
Sandstone, gray, friable	4	0
Shale, gray	3	7
Sandstone, argillaceous	1	4
Shale, drab, hard	1	2
Lignite	4	5
Shale, dark, hard	2	10
Shale, light, fossils, leaves	2	10
Lignite	2	6
Shale, carbonaceous	1	8
Sandstone	2	8
Shale	1	8
Sandstone and shale, alternately	2	8
Lignite	1	0
Sandstone and shale	3	0
Lignite		6
Sandstone	4	0
Lignite	1	7
Sandstone, white, hard	3	0
Sandstone in upper part, shale in lower part; *Ostrea glabra* above base (only representative found of Cannonball marine member)	33	11

[1] See footnote, p. 533.

	Ft.	In.
Lignite	1	11
Sandstone and shale, sandstone in upper part and shale below	27	0
Lignite		10
Shale	11	0
Lignite	3	0
Shale	3	4
Sandstone, hard ledge in upper part	13	2
Lignite	5	3
Shale	1	8
Lignite	1	2
Shale	1	4
Lignite	5	0
Shale	8	6
Lignite	6	11
Shale	4	8
Lignite		6
Shale	7	6
Lignite	5	4
Sandstone	15	4
Shale		
	250	11

SECTION ALONG NORTH FORK OF GRAND RIVER AT BLOOM, SOUTH DAKOTA[1]

	Ft.	In.
Fort Union formation(?)		
1. Sandstone, brown and yellow, fine-grained, thin-bedded, interbedded with lenses of compact, bluish-gray limestone		
Cannonball marine member of the Lance formation		
2. Sandstone, dark gray, very calcareous, marine fossils *Nucula* sp., *Callista deweyi*, *Pholas* sp., *Corbula* sp., *Anchura americana*	10	0
Ludlow lignitic member of the Lance formation		
3. Lignite	2	0
4. Shale and sandstone, interbedded	40	0
5. Lignite	1	0
	53	0

NOTE.—No. 1 was considered as Fort Union but very possibly is Cannonball, as indicated by more recent field work in other parts of the field.

[1] See footnote, p. 533.

GENERALIZED SECTION ALONG NORTH FORK OF GRAND RIVER BETWEEN BLOOM, SOUTH DAKOTA, AND HALEY, NORTH DAKOTA[1]

	Ft.	In.
1. Sandstone, brown and yellow, fine-grained, thin-bedded, and lenses of limestone	30	0
2. Sandstone, marine; burrows of a Teredo-like shell, *Anchura americana*	10	0
3. Shale	1	0
4. Lignite	2	0
5. Shale and sandstone, interbedded. Sandstone in lower part, light brown, medium fine-grained, micaceous; containing *Halymenites major*, *Thracia* sp., *Nucula* sp., *Callista* sp., *Pholas* sp., imprints of pelecypods and shark teeth	40	0
6. Lignite	1	0
7. Shale	4	0
8. Lignite	2	3
9. Shale		
	90	3

NOTE.—No. 1 was considered Fort Union but may be Cannonball; Nos. 2 and part of 5 are marine, and Nos. 3, 4, 6, 7, 8, and 9 are considered fresh-water.

Cannonball marine member.—The Cannonball marine member of the Lance formation has been traced and mapped in an area extending from near Mandan, North Dakota, to the vicinity of Ralph post-office in Harding County, South Dakota, and 4 miles west of Haley, Bowman County, North Dakota. The presence of brackish-water fossils, *Ostrea glabra*, near Yule on Little Missouri River in Billings County, North Dakota, shows that the sea probably extended some distance farther to the west than its sediments have been mapped. The extent of this member east of Missouri River is unknown.

The marine member is composed predominantly of dark sandy shale or shaly sandstone with a subordinate amount of dark-yellow and gray sandstone. It also contains some thin limestones. All the strata are lenticular in character and individual beds can be followed for only short distances. The member is typically exposed in the bluffs of Cannonball River in Tps. 132 and 133 N., Rs. 87 and 88 W. The following sections measured in the type area show

[1] The last three sections expose the interrelations of the Ludlow lignitic and the Cannonball marine members.

the character of the strata of the Cannonball marine member and of the strata which underlie and overlie it.

SECTION NEAR BASE OF THE CANNONBALL MARINE MEMBER, IN SOUTH BLUFF OF CANNONBALL RIVER, SEC. 33, TP. 132 N., R. 87 W.

	Ft.	In.
1. Sandstone and sandy shale, yellowish gray	23	0
2. Sandstone, yellowish gray, unconsolidated, with thin lenses of shale near the bottom, containing poorly preserved leaf impressions	22	0
3. Shale, brown, sandy	22	0
4. Shale, bluish brown	2	0
5. Lignite		4
6. Shale, bluish brown	4	0
7. Sandstone, gray, unconsolidated	45+	0
8. Concealed to river level	15	0
	133	4+

In this section no definite line could be drawn between the lower part of the Lance and the Cannonball member. No. 1 and probably No. 2 belong to the marine member.

SECTION IN SOUTH BLUFF OF CANNONBALL RIVER, ABOUT THREE-FOURTHS OF A MILE WEST OF FOREGOING SECTION

	Ft.	In.
1. Sandstone, yellow	25	0
2. Shale, containing poorly preserved plant remains	4	0
3. Sandstone, gray		
	29	0

The sandstone No. 1 in this section appears to be unconformable on the shale. It is probably the lowest bed of the Cannonball member.

SECTION IN BLUFFS OF CANNONBALL RIVER, NEAR JANESBURG, NORTH DAKOTA, TP. 132 N., R. 88 W.

	Ft.	In.
1. Sandstone, fine-grained, containing "cannonball" concretions	40±	0
2. Shale, dark, sandy	10	0
3. Sandstone and sandy shale, light colored, with "cannonball" concretions	21	6
4. Shale, dark, sandy, with *Perna* n.sp.	4	0

	Ft.	In.
5. Sandstone and shale, light colored, thin-bedded, the shale containing indeterminable vegetable remains	30	0
6. Sandstone, consolidated	1	6
7. Shale, light blue	1	6
8. Sandstone and shale, alternating	15	0
9. Shale, dark, carbonaceous	1	0
10. Sandstone and shale, alternating, base concealed by talus at base of slope	5+	
	129	6+

No. 4 and the overlying beds of this section are undoubtedly of marine origin; the lowest beds may be non-marine but no definite line could be drawn.

SECTION OF PART OF CANNONBALL MARINE MEMBER IN SOUTH BLUFFS OF CANNONBALL RIVER, E. ½ SEC. 5, TP. 132 N., R. 88 W.

	Ft.	In.
Shale, sandy, black and brown, with *Pyrifusus newberryi* (?), fragments of *Crassitellites*, and a coral belonging to an undetermined genus	16	6
Sandstone, yellow, largely consolidated, but with lenticular concretions of hard sandstone	31	6
Shale, sandy, dark gray, nearly black	52	0
	100	0

SECTION OF LOWER PART OF FORT UNION FORMATION AND UPPER PART OF CANNONBALL MARINE MEMBER OF LANCE FORMATION IN SOUTH BLUFF OF CANNONBALL RIVER, SECS. 31 AND 32, TP. 133 N., R. 88 W.

	Ft.	In.
Sandstone, yellow. Upper part massive, lower part thin-bedded and near base interbedded with thin bands of gray shale	30	6
Sandstone, dark gray to black	3	6
Sandstone, yellow, thin-bedded, interbedded with gray shale	13	6
Sandstone, gray, concretionary		6
Shale, bluish gray and somber-sandy	52	0
Shale, dark gray	54	6
Sandstone, yellowish gray, unconsolidated	5	0
Sandstone, yellow, hard at top, containing *Lunatia concinna*, *Turris minor*, *Fasciolaria buccinoides*, *Pyrifusus* sp., *Anchura* sp., and an undescribed coral	6	0
Shale, sandy, dark gray, grading at top into sandstone	31	6
	197	0

The upper sandstone of this section probably belongs to the Fort Union formation and the remainder to the Cannonball member of the Lance formation.

Sections similar in general character to those given above are exposed in the bluffs of Heart River and on Cedar Creek. The general appearance of the more sandy rocks is very much like that of the Fox Hills sandstone. The shale is very much like the Pierre.

A peculiar feature of both the Fox Hills sandstone and the Cannonball member of the Lance is the abundance of round concretions commonly known as "cannonballs." They are formed by cementation of the sandy shale by the deposition of calcium carbonate. These are true septarian nodules with radiating and concentric veins of calcite. The best examples observed are in a railway cut a few miles west of Raleigh, North Dakota. Where the "cannonballs" are exposed by stream erosion they are mostly weathered and broken to pieces.

The rocks of the Cannonball member weather typically into rounded hills, and in the interstream areas natural rock exposures are very few. In a large part of the area where the member was examined in South Dakota the only evidence that the beds are of marine origin are a few thin beds of nodular fossiliferous limestone, the presence of which is shown by the lines of residual bowlders at the borders of level-topped hills. The topography is such, however, that these horizons can be followed for long distances. No definite line could be drawn in the field between the Cannonball marine member and the lower part of the Lance. The contact of the two groups of strata is exposed at only a few places and in all such cases it seems to be impossible to tell where the beds of non-marine origin stop and those of marine origin begin. It follows from the foregoing statement that there is no evidence of unconformity at this horizon.

The relationship of the Cannonball member with the underlying lower part of the Lance is well shown in the region northwest of Solen on Cannonball River, North Dakota. In this area the characteristic chaos of badlands, formed by the erosion of the sandstone and shale of the lower part of the Lance, is bordered on

the north by a high, nearly level plateau which is capped by the lower fossiliferous sandstone of the Cannonball member. The top of the plateau is approximately 500 feet above and only about 2 miles distant from the river. The characteristic badland-forming strata of the lower part of the Lance extend up to within about 100 feet of the top of the plateau but here, as elsewhere, no sharp line of demarkation can be drawn.

Dr. Stanton has kindly prepared the following statement concerning the fauna of the Cannonball member.

The fauna of the Cannonball marine member of the Lance may now be characterized as a modified Fox Hills fauna. It contains a considerable proportion of undescribed species of Cretaceous affinities, and it is noteworthy that a number of the most common Fox Hills species have not been discovered in this fauna. The list of forms recognized is as follows:

Nodosaria sp.
Caryophyllia? sp.
Anomia sp.
Perna sp.
Crenella sp.
*Cucullaea shumardi M. and H.
†Glycimeris subimbricata (M. and H.)
*Leda (Yoldia) scitula M. and H.
*Leda equilateralis M. and H.?
*Nucula planimarginata M. and H.
‡Crassatellites evansi (H. and M.)
Solemya? sp.
*‡Lucina occidentalis (Morton)
Corbicula cytheriformis M. and H.
†Cyprina ovata M. and H.
*Cyprina ovata var. compressa M. and H.?
Veniella? sp.
Callista sp. *a*
Callista sp. *b*
Tellina? sp.
Thracia sp. related to T. subgracilis Whitfield
†Teredo globosa M. and H.
†Teredo selliformis M. and H.
Corbula sp.
Entalis sp.
Scala? sp.
Turritella? sp.
‡*Lunatia concinna (H. and M.)
Cerithium? sp.
‡*Anchura americana (E. and S.)
Anchura americana (E. and S.) robust variety
Helicaulax? sp.
†*Cantharus (Cantharulus) vaughani M. and H.
*Pyrifusus (Neptunella) newberryi M. and H.?
*Fasciolaria buccinoides M. and H.
*Fasciolaria (Piestochilus) culbertsoni M. and H.
*Turris contortus M. and H.
Turris sp. related to T. contortus M. and H.
*Turris minor (E. and S.)?
Cinulia sp.
*Cylichna scitula M. and H.?

In this list of about 40 forms there are 21 named species and varieties of which 15 (marked *) occur in the Fox Hills, 4 (marked ‡) occur in the Pierre, and 5 (marked †) were originally described from rocks now known to belong to the marine member of the Lance. One species, *Corbicula cytheriformis*, was described from the Judith River formation and is known in the Mesaverde

formation and the Lance of other areas. The fossil seaweed, *Halymenites major*, which is common in the Fox Hills and other sandy formations of the marine Cretaceous, is also associated with the above listed fauna.

FORT UNION FORMATION

West of Missouri River Fort Union formation overlies the Cannonball member of the Lance and on Little Missouri it overlies the Ludlow lignitic member of the Lance. In this part of the Dakotas a large part of the formation has been removed by erosion, but in Billings County, North Dakota, one of the writers (Hares) has found a thickness of 1,025 feet. The formation consists of calcareous sandstone and shale of continental origin, containing numerous thick persistent beds of lignite and an abundant flora and fresh-water invertebrate fauna. Some of the thick lignite beds have burned extensively, and great numbers of red hills composed of fused and baked rock are characteristic features of the formation. The lower 100 feet of the formation is made up almost wholly of partially consolidated yellow and gray fine-grained sandstone which is in some localities indistinguishable from the sandstone at the top of the Cannonball member. At numerous other exposures, however, where the upper bed of the Cannonball member is a sandy shale, the contact is easily followed. At one locality on the north bank of Heart River in Tp. 136 N., R. 88 W., there is an erosion channel from 30 to 50 feet deep in the Cannonball member filled with the channel deposits of a Fort Union stream. Similar channel sandstones at the base of the Fort Union were seen along Little Missouri, near Yule, and at the mouth of Deep Creek, and also on Sand Creek, in Billings County, North Dakota. These strata show evidence of a rapid change in the character of the sedimentation. They consist predominantly of coarse sandstone containing lenses or pockets of conglomerate and of soft clay shale. The conglomerate consists of pebbles derived from the strata of the surrounding region and contain numerous water-worn bones, teeth, fish scales, fragments of silicified wood, and lignite in the form of tree trunks. Among the vertebrate remains are two mammalian teeth which have been identified by Dr. J. W. Gidley of the U.S. National Museum as *Euprotogonia* sp., second lower molar of left jaw, and *Pantolamda cavirictus*, upper premolar. Both of these species are found in the Fort Union beds of Sweetgrass

County, Montana. The remaining fragments are identified as representing *Champsosaurus*, a crocodile, a turtle, and *Lepisosteus*.

Lenses of conglomerate similar to the one described above were seen at other localities near the base of the Fort Union along Heart River, and in one exposure a bed of conglomerate a few inches thick forms the base of the formation. Along Little Missouri in sec. 31, Tp. 138 N., R. 102 W., a conglomerate at the base of the Fort Union contains bowlders up to a foot in diameter. The unconformity shown in these exposures is such as would be expected in a transition from marine to continental sedimentation.

ROCKS OF POST-FORT UNION AGE

The rocks of post-Fort Union age in this region embrace: (1) small remnants of sandstone, marl, and limestone of the White River formation on the tops of a few high buttes; (2) sand and gravel beds on the tops of high buttes, probably deposited by streams previous to the present cycle of erosion and derived in part from the White River formation; (3) terrace gravel in the valleys of the larger streams from 50 to 250 feet above the present valley floors; and (4) a large number of scattered glacial bowlders, the remnants of the drift of one of the earlier glacial epochs. These later rocks occupy small areas and are not shown on the accompanying map.

INTERRELATIONSHIPS OF THE LUDLOW LIGNITIC AND THE CANNONBALL MARINE MEMBERS OF THE LANCE FORMATION

It has been shown that the Lance formation in a large region immediately west of Missouri River consists of two parts, the upper of which, the Cannonball member, is marine and contains a fauna similar to, but not identical with, that of the Fox Hills sandstone. The Cannonball member becomes gradually thinner toward the west, and the sea in which it was deposited perhaps did not extend as far west as the Montana line. The oyster beds near Yule in Billings County, North Dakota, first discovered by Leonard and later described by Stanton,[1] may represent the western-

[1] T. W. Stanton, "The Age and Stratigraphic Relations of the 'Ceratops Beds' of Wyoming and Montana," *Washington Acad. Sci., Proc.*, XI, (1909), 249; "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota and Eastern Wyoming, *Am. Jour. Sci.*, 4th ser., XXX (1910), 183–84.

most limit to which the Cannonball sea extended. This region was studied in 1911, and it was found by Hares that the beds containing the oysters are about 700 feet above the base of the Lance.

The zone in which *Ostrea glabra* and *Ostrea subtrigonalis* occur, in Tps. 136, 134, and 134 N., R. 105 W., is considered the westward extension (so far as known at present) of the strata of marine origin, as the oysters are brackish-water animals and consequently must have had some connection with the open sea. The most westerly collection of the Cannonball marine fauna is only 30 miles east of this place, in sec. 21, Tp. 129 N., R. 100 W., 4 miles west of Haley, and occurs stratigraphically within 100 feet of the T Cross lignite bed which was traced from Tp. 134 N., R. 105 W., to the west side of Tp. 129 N., R. 101 W. The oysters occur about 70 feet above the same lignite bed; it is assumed that the seaward connection was to the east. The oysters also occur about 120 feet below the base of the Fort Union formation which in the Little Missouri region has the same characteristics (channel conglomerate, light-yellow, somewhat massive sandstone) as it does in the Cannonball River country where it rests directly on the Cannonball marine member. It appears that the sea in which were deposited some 300 feet of marine sediments transgressed westward across the lignitic strata of the Ludlow member and that the position of its westward limit is underneath the divide between the drainage of Little Missouri and that of Grand and Cannonball rivers. All of the Triceratops collected in the Little Missouri country came from below the T Cross lignite bed and the oysters from above it. Calvert, however, states that in Montana "Ceratopsian bones were found just above the lowest persistent lignite bed, but there is certainly nothing in the character of the overlying strata to suggest that similar bones do not occur therein up through a stratigraphic distance of perhaps 500 feet."[1] The T Cross lignite bed was mapped to the Montana state line and it is undoubtedly the same lignite as the "persistent lignite" referred to above.

The presence of numerous lignite beds in the upper part of the Ludlow lignitic member of the Lance is in strong contrast to the

[1] W. R. Calvert and Others, "Lignite in Eastern Montana," *U.S. Geol. Survey Bull. 471*, p. 197, 1912.

undifferentiated lower part of the formation, and the absence of marine fossils in this member is in contrast to the Cannonball marine member. The Ludlow lignitic and the Cannonball marine members are considered to be contemporaneous in age.

RELATIONSHIPS IN OTHER AREAS

The Cannonball sea presumably advanced into western North and South Dakota from the east or northeast, and by inference the Cannonball member continued with undiminished or with increased thickness to the north and northeast. Practically the whole region east of Missouri River is deeply covered with a mantle of the drift of the last Wisconsin ice invasion, and the underlying formations are exposed only in isolated localities. In this region the Lance formation has been mapped about 50 miles eastward from Bismark but is presumably unrecognizable farther northeast.[1] It seems a very reasonable supposition that the lower fresh-water part of the Lance may not extend very far to the east and that where it is absent marine sedimentation was continuous throughout the time when the fresh-water Lance was being deposited elsewhere.

Beyond the limits of Cannonball marine sedimentation on the west the Fort Union formation rests directly on the fresh-water beds of the Lance, and, except along the badlands of Little Missouri, no definite line of demarkation between the two has been drawn. In general the Lance is distinguished lithologically by a greater proportion of somber-colored shale and sandstone, the large number of the bog iron ore nodules and layers, by irregularity of deposition, and by a paucity of lignite. The Fort Union, on the other hand, contains a larger proportion of yellow and white sandstone, quartzite, and thin limestone, and almost everywhere contains a large number of lignite beds, some of which are very thick and persistent. The rocks of this formation are fine-grained and regularly bedded and give rise to angular topographic forms. In a large area in eastern Montana, mapped by the U.S. Geological Survey in 1910, it was found that the color change from somber below to yellow above does not take place everywhere at

[1] A. G. Leonard, "The Geological Map of North Dakota," *Quar. Jour. Univ. N.Dak.*, IV (1913), 4.

the same horizon. One example is cited by Calvert[1] where in 2 miles the somber beds transgress a distance of 200 feet upward into the yellow.

At present the only accepted criterion for the distinction of the Lance formation from the Fort Union in areas where the Lebo andesitic member of the Fort Union[2] and the Cannonball marine member of the Lance are absent is the presence in the former of Triceratops and its associated fauna. On the basis of its fauna and flora the Lance formation is approximately correlated with the Denver and Arapahoe formations of the Denver Basin, Colorado,[3] and with a part of the "Upper Laramie" described by Veatch[4] in Carbon County, Wyoming. In both these regions the supposed Lance equivalents usually rest unconformably upon a thick series of fresh- and brackish-water beds—the Laramie of the Denver Basin and the "Lower Laramie" of Carbon County, Wyoming. Thus the correlations made on the basis of the vertebrate fauna and the flora indicate that the Laramie is older than the Lance. In the Denver Basin the Laramie rests conformably on the Fox Hills sandstone.

A comparison of the stratigraphic section in the Denver Basin and adjoining regions with that in North and South Dakota presents certain anomalies which are not easy of interpretation. Knowlton[5] has maintained that there is an unconformity throughout the northeastern Plains region between the Fox Hills sandstone and the Lance which represents the time equivalent of the whole of the Laramie of the Denver Basin as well as an unconformity

[1] W. R. Calvert, "Geology of Certain Lignite Fields in Eastern Montana," *U.S. Geol. Survey Bull. 471*, p. 197, 1912.

[2] G. S. Rogers, "The Little Sheep Mountain Coal Field, Dawson, Custer, and Rosebud Counties, Montana," *U.S. Geol. Survey Bull. 531*, pp. 168–172, 1913.

[3] S. F. Emmons, Whitman Cross, and G. H. Eldridge, "Geology of the Denver Basin in Colorado," *U.S. Geol. Survey Mon. 27*, 1896.

[4] A. C. Veatch, "On the Origin and Definition of the Geologic Term 'Laramie,'" *Am. Jour. Sci.*, 4th ser., XXIV (1907), 18–22; *Jour. Geol.*, XV (1907), 526–49.

[5] F. H. Knowlton, "The Stratigraphic Relations and Paleontology of the 'Hell Creek Beds,' 'Ceratops Beds,' and Equivalents, and Their Reference to the Fort Union Formation," *Washington Acad. Sci., Proc.*, XI (1909), 179–238; "Further Data on the Stratigraphic Position of the Lance Formation ('Ceratops Beds')," *Jour. Geol.*, XIX (1911), 358–76.

above the Laramie. Numerous localities in Wyoming, Montana, and North and South Dakota are described by Knowlton where this unconformity is supposed to be shown, (1) by erosion channels in the Fox Hills sandstone; (2) by discordance of dips between the two formations; or (3) by the great variation in thickness of the Fox Hills sandstone. In each case, however, the facts can be explained on a theory of practically continuous and unequal deposition.[1] Furthermore, the presence in the Dakotas of a marine fauna very similar to that of the Fox Hills sandstone, overlying the fresh-water sediments of the Lance, renders less tenable the theory of an unconformity of any importance at the base of the Lance, since the open sea must have persisted throughout Lance time in a region not very remote from western North Dakota.

Two other explanations have been suggested, the first that the fresh-water Lance strata of the Dakotas are the equivalents of the Laramie, Arapahoe, and Denver formations of the Denver Basin and of both the "Upper Laramie" and "Lower Laramie" of Carbon County, Wyoming. In this case the unconformity at the base of the Arapahoe is not of widespread general importance. The other and, to the mind of the writers, the more plausible hypothesis is that the Fox Hills sandstone of the Denver Basin is not the time equivalent of that in the type locality in South Dakota but is considerably older. The Fox Hills sandstone is, under this supposition, merely a near-shore phase of the sedimentation of the Montana sea. This sea persisted to much later time in the Dakotas than in Colorado, and the sedimentation was practically continuous until the close of the Fort Union. The possibility of finding a region where sedimentation was continuous throughout Laramie, Arapahoe, and Denver time was pointed out by Cross in 1909.[2] This deposition, it would seem, continued in the Dakotas in the open sea until near the close of Laramie and probably also throughout the period of erosion following the Laramie. Then the sea

[1] T. W. Stanton, "Fox Hills Sandstone and Lance Formation ('Ceratops Beds') in South Dakota and Eastern Wyoming," *Am. Jour. Sci.*, 4th ser., XXX (1910), 172–88.

[2] Whitman Cross, "The Laramie Formation and the Shoshone Group," *Washington Acad. Sci., Proc.*, XI (1909), 35.

withdrew still farther to the east and with the beginning of Lance time continental sedimentation was inaugurated over a large part of the eastern Rocky Mountain and Great Plains region. The last advance of the sea into the region of the Dakotas was approximately contemporaneous with the final dying-out of the Cretaceous dinosaur fauna, but there was no marked break in the continuity of sedimentation until the close of the Fort Union. Then the area suffered deformation and the erosion of thousands of feet of strata before the deposition of the later formations which rest unconformably on the Fort Union and all the underlying formations.

IS THE LANCE FORMATION TERTIARY OR CRETACEOUS?

In a general consideration of the age of the late Cretaceous or early Tertiary formations in the Rocky Mountain and Great Plains regions there are five lines of evidence which must be taken into consideration, and the decision at which any geologist will arrive will depend largely upon the relative importance which he gives to one or another of these lines of evidence. In a paper dealing primarily with the stratigraphy of a limited area in the Great Plains the larger and more general problem can be outlined only very briefly.

In the opinion of the writers the greatest weight should be given to the evidence of diastrophism. This, however, is of undetermined significance. The greatest break in the sedimentary record comes at the close of the Fort Union where there was everywhere, so far as known, in the Rocky Mountain and Great Plains regions an interval of long erosion and in many places of extensive folding. An earlier but apparently much less extensive break in the continuity of sedimentation is represented by erosional unconformities of greater or less extent at the base of the Raton, Dawson, Arapahoe, and "Upper Laramie" formations in New Mexico, Colorado, and Wyoming. An attempt to show that this unconformity is general throughout the northern plains region has, as previously pointed out, met with little success. In the Gulf and Atlantic Coastal Plains provinces the interval between the Cretaceous and Tertiary systems is marked by a great change in marine faunas and presumably by a profound break in the sedimentary

record. Obviously, if a definite correlation can be made of this unconformity in the Coastal Plain region with an unconformity in the Rocky Mountain region a long step has been taken toward settling the problem for all of the western region.

This leads directly to a second line of evidence—the fossil flora. The recent work of F. H. Knowlton[1] and E. W. Berry has shown a close relationship between the floras of the Midway and Wilcox formations of the Gulf region, which is indisputably of age, with the floras of the Raton and the Denver formations Eocene of New Mexico and Colorado. This evidence, together with the fact that the Raton and the Denver and Arapahoe formations rest unconformably on underlying formations which contain totally different fossil floras, has led Knowlton to the conclusion that the Raton and Denver formations are of Tertiary age. This conclusion is now accepted by the U.S. Geological Survey, but some geologists question the correlation, as it depends mainly on the evidence of the fossil flora.

When we turn to the evidence gathered from a study of the marine invertebrates we arrive at a diametrically opposite conclusion. As has been pointed out in preceding pages, the marine fauna of the Lance formation is distinctly a Cretaceous fauna very closely related to the fauna of the Fox Hills sandstone. It has been pointed out too that the Lance and Fort Union formations should be grouped together both on account of the close relationship of their fossil floras and because in places no lithologic separation of the two can be made. Thus the presence of a marine Cretaceous fauna in the Lance indicates that the Lance and probably also the Fort Union should be placed in the Cretaceous system. This would make the line between the Cretaceous and Tertiary correspond with the major unconformity above the Fort Union. It must be admitted that in the Rocky Mountain and the Great Plains regions no marine fauna of undoubted Tertiary age is known with which the Lance fauna may be compared, and there is a possibility that in this area the Fox Hills marine

[1] F. H. Knowlton, "Results of a Paleobotanical Study of the Coal-bearing Rocks of the Raton Mesa Region of Colorado and New Mexico," *Am. Jour. Sci.*, 4th ser., XXV (1913), 526–30.

fauna lived over from the Cretaceous into the Tertiary and thus became contemporaneous with the totally different marine faunas of the Gulf and Atlantic coasts, but if this did happen it is the first recorded case of such contemporaneous Cretaceous and Eocene faunas.

The evidence of the reptilian fossils in the Lance formation supports the evidence of the marine fauna. This—the Triceratops fauna—is the last of the great dinosaur faunas of the Mesozoic and for that reason has been considered generally by vertebrate paleontologists as sufficient evidence of the Mesozoic age of the formation. It is, however, just as probable that the Ceratopsian dinosaurs lived over into the Tertiary as that the marine Fox Hills fauna did so.

There remains yet to be considered the evidence of the mammalian fauna of the Fort Union formation. This fauna, of course, has its closest affinities with the succeeding Eocene faunas, since practically no mammalian fauna is known in the later Cretaceous formations. There is here the possibility that the Tertiary mammalian faunas began their development before the close of the Cretaceous. Lee[1] has recently described a find of a bone of a Creodont mammal associated with Ceratopsian bones in the Dawson arkose which shows that Ceratopsian dinosaurs and mammals of Tertiary aspect must have lived at the same time.

Such in outline is the Lance problem as it stands today. A recent detailed consideration of all of the evidence has led to a decision by the U.S. Geological Survey that the Denver and Arapahoe, Dawson, and Raton formations in Colorado and New Mexico all be placed in the Tertiary system. This decision was based primarily on the correlation of these formations with the Wilcox formation of the Gulf region on the evidence of their fossil floras and also on the consequent correlation of the unconformities in the two regions. Although the Lance formation is believed to be of about the same age as the Denver, Raton, and "Upper Laramie," it is classified by the U.S. Geological Survey

[1] W. T. Lee, "Recent Discovery of Dinosaurs in the Tertiary," *Am. Jour. Sci.*, 4th ser., XXXV (1913), 531–34.

as Tertiary(?), the doubt being thus expressed on account of the Cretaceous character of the Cannonball marine fauna. The writers believe that greater weight should be given to the evidence of the marine faunas and the dinosaurs, and that, in view of the strong evidence presented by these faunas, the correlations made on the basis of fossil floras should not be considered as conclusive.[1]

[1] This paper was submitted for publication in the *Journal of Geology* in June, 1914. Further studies on the Lance problem have strengthened the conviction of the writers that the Lance is Cretaceous.

VARIATIONS OF GLACIERS. XIX[1]

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The following is a summary of the *Eighteenth Annual Report* of the International Committee on Glaciers.[2]

THE REPORT OF GLACIERS FOR 1912

Swiss Alps.—The tendency of the Swiss glaciers to advance, which began in 1910, and which disappeared in 1911, following the wonderfully clear summer of that year, again became evident in 1912 (and more definitely than in 1910), probably on account of the very cold and wet summer of 1912. Although the number of glaciers retreating is still larger than the number of those advancing the former number is diminishing and the latter increasing.

Eastern Alps.—The heavy snowfall, the low summer temperature, and the hazy condition of the atmosphere reduced the melting during the summer to such an extent that even in August the glaciers were still covered with snow to their ends. One would expect, as a consequence, a marked advance of the glaciers, but this was by no means the case; of the thirty-four glaciers observed, fully twenty-eight were clearly in retreat, three were stationary and only three (and they were small) were advancing. Whether the advance of these three was due to the conditions holding during this one summer, or whether it is the forerunner of a general advance must be determined by the future. We can summarize by saying that the retreat of the glaciers of the Eastern Alps continues, but to a less marked degree than last year.

Italian Alps.—Observations were made in the Piedmont, the Lombard, and the Venetian Alps; they were greatly interfered with by the heavy snowfall, but the glaciers still continue to retreat.

French Alps and Pyrenees.—The tendency of the glaciers to advance, shown in 1911, has not been maintained in 1912. Some

[1] Earlier reports appeared in the *Journal of Geology*, III–XXI.

[2] *Zeitschrift für Gletscherkunde*, VIII (1913), 42–62.

glaciers show slight advances, others slight retreats. The snowfall was very heavy, but it has not yet affected the extent of the glaciers. In the basin of the Gave de Pau, Pyrenees, the snowfall in the winter of 1911–12 was small, but on account of the small summer melting very many of the glaciers remained covered with snow at the end of the summer. The glaciers increased in thickness especially in their reservoirs. On the whole, the glaciers observed have shown for several years a definite tendency to advance.

Swedish Alps.—The two glaciers observed show some slight advance.

Norwegian Alps.—All the glaciers of the Jotunheim are retreating. In the Folgefon and Jostedalsbrae, along the west coast, eighteen glaciers are retreating and six are advancing. In the more northerly regions eight are retreating and three are advancing. A larger proportion of the glaciers under observation were retreating in 1911 than in 1912.

Greenland.—A small tongue of the inland ice near Disco Bay experienced an advance which culminated in July, 1912. Later in the summer the ice had retreated. The large tongue in the same neighborhood, called the Ekip Sermia, seemed to be stationary. The number of icebergs in the Jacobshavn Fiord was becoming smaller. Three small glaciers on Disco Island, first described by Chamberlin in 1894, and mapped in 1897, have since that date retreated between 30 and 80 meters. There are also other indications of the diminution of the glaciations. Recent moraines in front of several glaciers in Sermilik Fiord show that these glaciers are retreating.

REPORT OF THE GLACIERS OF THE UNITED STATES FOR 1913

The Arapahoe Glacier, in Colorado, shows no appreciable change (Henderson).

Professor Lawrence Martin sends me the following information regarding the variations of Alaskan glaciers in 1913:

Glacier Bay.—Several glaciers in this fiord were studied by the writer in 1913, under the auspices of the National Geographic Society. Grand Pacific Glacier, which retreated 7,425 feet between June 1 to August 1, 1912, had advanced 4,000 feet by September 9, 1913, so that it again terminates south

of the International Boundary. De Margerie Glacier and the Reid and Lamplugh glaciers at the upper end of Glacier Bay were likewise active and advancing, the two latter moving forward half a mile from 1911 to 1913. Rendu Glacier retreated about 4,100 feet from 1911 to 1913. The adjacent cascading glacier, which was advancing in 1911, was retreating, and no longer reached tide water in 1913. The southwesternmost tributary of Rendu Glacier advanced in 1913 and pressed out the medial moraines of the Rendu. A cascading glacier on the east was probably also advancing.

Adams Glacier in Muir Inlet retreated about 3,575 feet from 1907 to 1913. Muir also continued to recede. It was more accurately mapped in 1913 than at any time since 1892. Its total recession between these dates was about $7\frac{3}{4}$ miles. Its decrease in thickness is notable; according to Reid's Survey, the ice surface was 1,500 feet above sea-level in 1892 at the site of the ice-front of 1913; and nunataks, 1,050 and 1,150 feet high, were, at the earlier date, covered by ice to depths of 400–600 feet. Soundings in Muir Inlet in 1913 revealed a depth of 1,128 feet at a point where the ice surface was 1,250 feet above sea-level in 1892. Thus the total thickness of Muir Glacier $5\frac{5}{8}$ miles from its 1892 terminus is now known to have been 2,378 feet. Soundings likewise show that in Tarr Inlet (at the upper end of Glacier Bay) the ice of Grand Pacific Glacier at a point 12 miles from the terminus of 1894 was at least 2,500 feet thick.

Near Lynn Canal.—R. G. McConnell reports that the Rainy Hollow or Sullivan Glacier had advanced about 2,000 feet before it was observed in 1910 by Webster Brown, making a total advance of about three-quarters of a mile in less than ten months. It receded slightly between 1910 and 1913. Jarvis Creek Glacier likewise advanced in 1910. The detailed map of the Mendenhall, Eagle, Herbert, and adjacent glaciers east of Lynn Canal,[1] is accompanied by cursory notes on these ice tongues by Knopf.

Yakutat Bay.—Since 1910 Hidden Glacier has retreated 400–500 feet at the terminus and thinned 150 feet or more at the north margin. The tidewater terminus of Nunatak Glacier retreated slightly from 1912 to 1913. The adjacent cascading glacier advanced several hundred feet from 1910 to 1913. Hubbard Glacier receded slightly. Turner Glacier suffered a more pronounced diminution, especially at its south margin.

Mount St. Elias Region.—North of Mount St. Elias the Logan-Chitina Glacier changed from spasmodic, earthquake-stimulated activity in 1912 to stagnation in 1913, as reported by D. W. Eaton and J. D. Craig of the International Boundary Surveys. Its margin receded 30–50 feet during the year. A number of the ice tongues south of Logan Glacier were first mapped in 1913. The surveyors saw no unusual activity between Logan Glacier and Mount St. Elias but report that ablation during cloudy, rainy weather in July lowered the ice surface at a measured rate of between 3 and 4 inches per day.

[1] *Bull. No. 502, U.S. Geol. Survey*, 1912.

The new Icy Bay at the western border of Malaspina Glacier was mapped in 1913 by A. G. Maddren of the United States Geological Survey. The recession of Guyot Lobe now amounts to nine or ten miles. Maddren's map indicates that the Bering Glacier, which has hitherto been considered a single piedmont ice mass, is really two separate piedmont bulbs (Coast Survey Chart 8002, edition of March 12, 1912). A similar condition at Grand Plateau Glacier was noted from close inshore by the special steamer of the International Geological Congress of 1913. In both cases the earlier mapping was in error, for there has been no recession of either the Grand Plateau Glacier or the Bering Glacier in historical times.

Copper River Canyon.—The Childs Glacier has retreated slightly and its northern margin, as measured by Caleb Corser, is now 1,500 feet from the railway bridge on the Copper River & Northwestern Railway. In spite of an advance of more than 60 feet in the interval, its net change, from June 16, 1911, to July 12, 1913, was a recession of 26 feet, and from the latter date to November 7, 1913, it retreated between 100 and 150 feet more. Allen Glacier, after a slight advance in 1912, began to recede in 1913. Between August 26, 1910, and July 12, 1913, it advanced 2650 feet, or four-tenths of the distance to the railway north of it. From the latter date to October 28, 1913, Corser's measurements show that its northwestern border retreated 300–700 feet. Schwan Glacier is said by a prospector to have advanced in 1912.

Wrangell Mountains.—In connection with her successful ascent of Mount Blackburn in 1912,[1] Miss Dora Keen made important observations of snowfall and temperature, giving an idea of the precipitation that supports the large Kennicott Glacier. The total snowfall for thirty-three days (April 22 to May 24) was 40 feet (packed to 20 feet). During this period the temperature at altitudes of from 2,000 to 16,140 feet above sea-level ranged from −6° F. to +70° F. At her base camp, altitude 5,500 feet, three feet of snow fell in thirteen consecutive days, while at a camp 6,900 feet higher the snowfall during the same period, which included four days of hot sun, was very much greater. It was measured by the burial of bamboo poles 12 feet long in front of a shelter excavated in the snow and may be estimated at 30 feet (packed to 10 feet). At the summit of Mount Blackburn the snowfall seems to be less, though it may be all carried away by the winds. An important part of the alimentation of Kennicott Glacier comes in great snow avalanches from the steep walls of its valley. Some of these avalanches, which are most abundant between noon and 4:00 P.M., advance three miles from the valley wall, or the whole width of the glacier.

Prince William Sound.—Barry Glacier in Harriman Fiord retreated an amount varying from about 2,500 feet on the western side to at least 7,200 feet on the eastern border between Martin's visit in July, 1910, and Johnson's in 1913. Coxe Glacier, formerly a tributary of the Barry is now independent.

[1] *World's Work*, XXVII (1913), 80–101.

Cook Inlet.—A glacier at the head of Tuxedni or Snug Harbor seems to have advanced between 1904 and 1911. Coast Survey Chart 8554 shows some of the glaciers on Mount Iliamna for the first time and new details of the glaciers on Mount Douglas.

In 1913 Archdeacon Stuck found that a marked change had taken place in one of the great ridges of Mount McKinley since Parker's ascent in 1912; the snow and ice had broken away from a larger portion of the ridge, owing, possibly, to the earthquake of July 6, 1912.

In the absence of M. Charles Rabot, the retiring president of the Commission International des Glaciers, his report to the Twelfth International Congress of Geologists at Toronto was presented by Mr. Emile de Margerie.[1] It may be summarized as follows:

At present the glaciers throughout the world are in general retreating. Though in some regions, as Iceland, the retreat is slight, in others, such as Alaska, Norway, and the Alps, it is very marked. Thus in the Pelvoux *massif*, several small glaciers have entirely disappeared in the last thirty years.

The retreat, which has lasted for a century in Norway, and for fifty or sixty years in the Alps has been interrupted by small temporary advances. In Norway an advance in the last few years is dying out; in the French Alps and in the Pyrenees a secondary advance is developing.

A careful search of old documents has revealed the variations of the Chamonix glaciers since 1580. A strong advance occurred during the last years of the sixteenth century; others in 1643, in 1663, and in 1716; the last culminated only about 1741. The variations since then were already known: a new advance began near the end of the eighteenth century, reached its maximum in 1818–20, and continued, but in a mild form, until the middle of the nineteenth century; since then the glaciers have been in retreat. It seems that the advance about 1600 was greater than had occurred for several generations before that date; and that, since then, the glaciers have not been as reduced as they are at present. In Norway, also, branches of the Jostedalsbrae and of the Svartisen advanced strongly in the first quarter of the eighteenth century; and in Iceland an increase of the glaciation seems to have existed

[1] The report is printed in the *Comptes rendus* of the Congress, pp. 144–48, and in *Zeitschrift für Gletscherkunde*, VIII (1914), 263–69.

since the tenth century. These records lead to the conclusion that glaciers are subject to variations of two orders: primary variations of great amplitude and of unknown duration; and, superposed on them, secondary variations, which, in the Alps, appear to follow Brückner's law. These oscillations are due to climatic variations, and it is advisable that brief reports of the snowfall should accompany the records of glacier variations.

Since the last congress the commission has suffered severely by the deaths of F. A. Forel, who may be looked upon as the originator of the study of glacier variations; of K. J. V. Steenstrup, who has done such excellent work in Greenland; and of R. S. Tarr, well known for his studies in Alaska. M. Paul L. Mercanton, of Lausanne, succeeds M. Forel as the ordinary member for Switzerland; M. Paul Harder takes the place of M. Steenstrup representing Denmark; and Mr. Alan G. Ogilvie now represents Great Britain in the place of Mr. Freshfield, who has retired and become a corresponding member. Other corresponding members elected recently are: Messrs. J. Rekstad, Adolf Holl, A. O. Wheeler, Lawrence Martin, and F. E. Matthes.

The officers of the commission elected to serve until the next meeting of the International Congress of Geologists are: Honorary President, Prince Roland Bonaparte; Active President, Dr. Axel Hamberg; Secretary, Dr. Paul L. Mercanton.

A NEW GENUS AND SPECIES OF AMERICAN THEROMORPHA

MYCTEROSAURUS LONGICEPS

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The past summer, Mr. Herman Douthitt, of the University of Chicago paleontological expedition, discovered on Mitchell Creek, Texas, in the horizon that has yielded various specimens of *Pantylus* and the type specimens of *Broiliellus* and *Glaucosaurus*, a broken skull and parts of the skeleton of a small reptile which, at the time, were thought to belong to the genus *Varanops*. Although it resembles that genus in shape and general characters, a more careful examination disclosed a new and peculiar form of Theromorpha.

The specimen is of a nearly white color, inclosed in a rather hard, siliceous red clay, from which the bone readily separates, leaving impressions as though made in wax. The skull had suffered a little from compression, but is otherwise undistorted. It was originally quite complete, but, as found, some of the bones had separated from the matrix and been lost; and the tip of the premaxillæ was gone. This partial loss of the bony structure, on one side or the other, while detracting from the appearance of the specimen, has made vividly apparent nearly every suture; others are clearly shown in the bone itself. The sutures of the inner side of the cranial bones are not always quite the same as on the outside; as an instance of which, the shape and size of the postfrontal of *Pantylus* is quite as Case and Huene figured it, while on the outside Broom gave the form correctly; and the quadratojugal of the same genus is nearly twice as broad on the inside as on the outside. In the figures given herewith (Figs. 1, 2) I have widened the face a little in top view, perhaps not quite enough. I have depended in no instance on any line indicating a suture unless it is precisely corroborated on the two sides, precluding the danger of mistaking cracks for

sutures. These figures will, I trust, render a detailed description for the most part unnecessary. The points of interest meriting

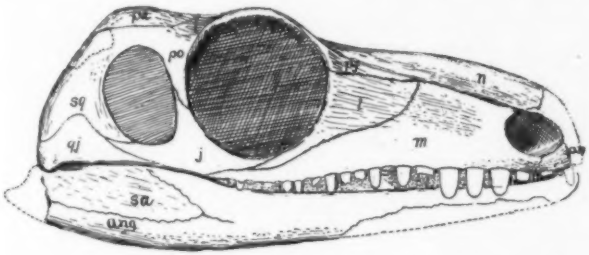

FIG. 1.—*Mycterosaurus longiceps* Will. Skull from the side. *n*, nasal; *pm*, premaxilla; *m*, maxilla; *l*, lacrimal; *pf*, prefrontal; *pof*, postfrontal; *po*, postorbital; *j*, jugal; *qj*, quadratojugal; *sq*, squamosal; *pa*, parietal; *f*, frontal; *ds*, dermosupraoccipital; *so*, supraoccipital; *sa*, surangular; *ang*, angular.

discussion are the nares, orbits, lacrimals, the boundaries of the temporal opening, and the structure of the occiput.

The nares are unusually large, as are also the orbits. The latter are nearly circular in outline, and look almost directly outward. They project above the frontal region, which is concave transversely between them. Within the orbits the broad smooth upper surface of the palatines is apparent, as also the narrow pterygoid; both bear small, conical teeth. The lacrimals do not reach the nares, as has been assumed without positive proof for some of the American Theromorpha. On the two sides not only do the sutural lines agree exactly, but the smooth, clean surface in front shows no trace of a suture on either side.

The jugo-postorbital arch is slender; on the right side it is bent inward, separating at the suture; on the left side it is broken in two or three places. The quadratojugal is definitely shown on each side. Because of the squamous

FIG. 2.—*Mycterosaurus longiceps*. Skull, from above. Explanations as in Fig. 1. Both figures natural size.

overlap of the squamosal one cannot be quite sure of the course of the suture externally, but it descends no lower than I have figured it. Above, the squamoso-postorbital arch is shown very decisively on each side, but I cannot be sure of the suture separating the two bones, and have, therefore, omitted it. I find no definite evidence of a tabulare or supra-temporal in this region, though it is not at all improbable that one or both are present. Posteriorly the squamosal covers the quadrate broadly. The temporal opening is moderately large and oval in form. It is bounded, it will be seen, almost precisely as in *Sphenodon*. The parietal on each side is slightly separated from the squamoso-postorbital arch, indicating one way in which the upper temporal opening in *Sphenodon* may have arisen. The parietal foramen is situated almost at the extreme posterior end of the parietals, and very close to the dermosupraoccipitals. On each side the parietals are curved downward to near the top of the quadrate, but I find no evidence of a suture separating the extremity of the processes.

The occipital surface is but little injured and quite clean of matrix. On the upper part of the steeper declivity, between the descending processes of the parietals, there is a thin, concave bone on each side, lying against the back part of the parietals, and somewhat separated in the middle line above by a descending process of the same bones. Below, each bone diverges a little to fit into a small groove on the upper external surface of the supraoccipital. The suture between these bones and the supraoccipital on each side is very conspicuous. The bones extend outward to the extremity of the posterior parietal processes. There is a small but distinct post-temporal opening on each side between the supraoccipital, dermosupraoccipital, and squamosal.

On the large surface below the dermosupraoccipitals and the temporal fossae I can distinguish no sutures. In shape, the surface is very much like that of *Dimetrodon*, as shown in the accompanying figures (Figs. 3, 4) made from a viewpoint at right angles to the plane of the supraoccipital. The paroccipital process is separated from the upper, supraoccipital process in the same way; and their union with the quadrate is very similar in both genera. I am aware that this bone in *Dimetrodon* has been differentiated into the dermo-

supraoccipital and supraoccipital, in addition of course to the paroccipitals, exoccipitals, and petrosal, but I can find no evidence whatever for the inclusion of the first-mentioned bone in the complex. I have sectioned various specimens, and studied numerous others. If, then, these bones correspond—and they surely do—there are no dermosupraoccipitals in *Dimetrodon*, or they have been reduced to the merest vestige, as in the modern gavials. As will be seen by the figure (Fig. 4, *po*) the surface of the supraoccipital articulating with the parietals above is cartilaginous, corresponding to the condition found in the dinosaurs, crocodiles, lizards, etc. The exoccipitals in *Mycterosaurus* surround the foramen magnum and are loosely attached, as in *Dimetrodon*. From below, not much is visible. The descending process of the paroccipital for the stapes is visible from behind, but I do not discover either the stapes or the tympanic, both of which are present in *Dimetrodon* (Fig. 4, *t*, *st*). This last figure is introduced here anticipatory of a full discussion of the cranial anatomy of this and other genera, which will be published later.

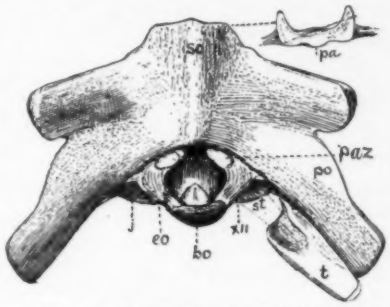

FIG. 3.—*Dimetrodon*. Occipital complex, obliquely from behind. *so*, supra-occipital; *paz*, proatlantal zygapophysis; *po*, paroccipital; *st*, stapes; *t*, tabulare; *bo*, basioccipital condyle; *eo*, exoccipital; half natural size.

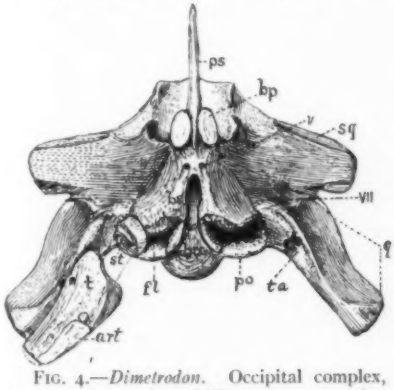

FIG. 4.—*Dimetrodon*. Occipital complex, view opposite to that of Fig. 1. *ps*, parasphenoid; *bp*, basipterygoid process; *bs*, basisphenoid; *bo*, basioccipital; *st*, head of stapes; *t*, tympanic; *ta*, articular pit in paroccipital for tympanic; *art*, articular surface; *po*, paroccipital.

The teeth in *Mycterosaurus* are characteristically different from those of *Varanops*, *Ophicaodon*, *Dimetrodon*, or *Edaphosaurus*. Altogether there are about eighteen, possibly one more, on each maxilla, the first four or five the largest. They are moderately elongated and flattened, with a rather obtuse apex.

Of the skeleton of this specimen, there are fragmentary vertebrae, and a coraco-scapula. The posterior coracoid is well ossified and not distinguishable from the anterior. It has a straight line below and is rather narrow. The interclavicle appears to be of the usual form, and very unlike the extraordinary interclavicle of *Pantylus*. The vertebrae, so far as they are preserved and prepared, are like those of *Varanops*. The spine is thin, and not more than two or three times the height of the centra.

Relationships.—There can scarcely be dissent from the opinion that *Mycterosaurus* is related to *Dimetrodon*, and should find its natural place in the same suborder, the Pelycosauria, though in a different family. The skull and vertebrae are more primitive, but both have the same relations of the bones, the same temporal opening, and the same elongated, narrow skull. The lower jaw lacks the inflected angular process below, but that is a specialization that would hardly be expected in the more primitive form. I am aware that this peculiarity has been largely relied upon as indicative of the relationships between the American and African Theromorpha, but I am skeptical of its value. No other American genus shows it, except, in a lesser degree, *Sphenacodon* and *Edaphosaurus*, and these are all highly specialized animals with elongated spines.

The family determination of *Mycterosaurus*, until more of the skeleton is known, is doubtful; it may provisionally be placed in the Poliosauridae. Of the American reptiles, nearly every specimen that can be differentiated by decisive characters represents a distinct genus; and the majority of genera, when fully known, are more or less justly placed in distinct families. The following genera, it seems to me, are rightly separated in different families; a fuller knowledge may require further division.

Sphenacodontidae
- Clepsydrops Cope
- Sphenacodon Marsh
- Dimetrodon Cope

Ophiacodontidae
  Ophiacodon Marsh
  Theropleura Cope
Poliosauridae
  Poliosaurus Case
  Varanosaurus Broili
  Varanops Williston
  ?Mycterosaurus Williston
  ?Arribasaurus Williston
Edaphosauridae
  Edaphosaurus Cope
Caseidae
  Casea Williston

Together with *Trichasaurus* Williston, *Scoliomus* Williston and Case, *Poecilospondylus* Case, *Glaucosaurus* Williston, *Elcobresaurus* Case, of more doubtful position, these are all that can be definitely located among the American Theromorpha, that is zygocrotaphous reptiles, with the temporal vacuity below the squamoso-postorbital arch. *Ostodolepis* Williston is a synonym of *Pantylus* Cope.

USE OF THE SLIDE RULE IN THE COMPUTATION OF ROCK ANALYSES

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Certain phases of petrographic and metamorphic work require the conversion of rock analyses into the corresponding approximate mineralogical compositions. By ordinary methods of calculation this may be rather prolonged and call for a considerable expenditure of time and energy. Exact conversion is, of course, impossible, especially without any preliminary data as to the minerals actually present, and, in the case of such variables as the biotites, amphiboles, and pyroxenes[1] their chemical composition itself should be determined also. Nevertheless the results attainable are of sufficient value to justify the procedure.

Various methods have been suggested, each possessing more or less merit and marking some advance in this direction. The use of molecular ratios has been very general and has been described by Kemp,[2] Cross,[3] and Osann.[4] Harker[5] calculated decimal tables for the different minerals which very materially assist in converting oxide values into the corresponding mineral values, but which are not adapted to the reverse process which in some cases is of equal importance. Mead[6] devised a set of graphic tables, one for each mineral, by means of which the value of any radicle in a mineral might be converted into the corresponding value of any other radicle

[1] Cross, Iddings, Pirsson, and Washington, *Quantitative Classification of Igneous Rocks*, 1903, p. 145.

[2] J. F. Kemp, "The Recalculation of the Chemical Analyses of Rocks," *School of Mines Quarterly*, XXII (1900–1901), 75.

[3] Cross, Iddings, *et. al.*, *op. cit.*, Part III, "Methods of Calculation."

[4] A. Osann, *Beiträge zur chemischen Petrographie*, I. Teil. "Molekular-Quotienten zur Berechnung von Gesteinsanalysen," Stuttgart, 1903.

[5] A. Harker, private publication, Cambridge University Press, 1910.

[6] W. J. Mead, "Some Geological Short-Cuts," *Economic Geology*, VII, No. 2 (March, 1912), 136.

in that same mineral or to the total mineral value. Conversely, the value of any radicle could be determined from the total mineral value. This marked a decided step in advance, but as a complete table is necessary for each mineral the disadvantages are obvious.

Later in the same article[1] Mead describes a "geologist's slide rule," which is a much more efficient method of conversion, and which has certainly merited a wider consideration than it has apparently received. He has devised a rule for ordinary rock minerals and another one for ore minerals, comprising a total of nearly 100 minerals, whereby very rapid conversions and comparisons are possible, either from constituent to mineral or the reverse. The rules are of white celluloid, about ten inches in diameter and are exceedingly simple to manipulate. In using these rules certain features became apparent to the writer: (1) a rule is good only for the minerals printed thereon; (2) ordinary slide-rule computations are impossible on the logarithmic scale of the rule; (3) the rule is too large for ordinary pocket use and hence of limited value in the field. These statements are not intended to detract in any way from the value of the "geologist's slide rule," but led the writer to consider a way to broaden this excellent application of the logarithmic scale to geologic work.

In a recent publication Lindgren[2] gives a summary of the various methods of recalculation. The scheme most favored "consists of a co-ordinate system in which the abscissas represent the distance from the vein which may be taken as the origin of the acting solutions, and the ordinates represent the molecular ratios multiplied by 100, except for silica, for which the scale must be reduced to bring the diagram within convenient compass." This method is excellent for just such cases as cited,[3] but a diagram so constructed is limited to one definite zone or locality and does not have the range of broad comparisons and generalizations inherent in the straight line diagram.

In the conversions and comparisons of chemical analyses referred to, considerable multiplication and division are necessary

[1] *Ibid.*, p. 139.

[2] Waldemar Lindgren, *Mineral Deposits*, 1913. Chap. xxx.

[3] F. L. Ransome, *U.S. Geol. Survey Professional Paper 75.*

in addition to that obviated by the "geologist's slide rule." For this work slide-rule accuracy is sufficient. In extended studies of this nature the use of the rule may save from 50 to 90 per cent of the time required for such calculations. Hence, if these conversions could be simply accomplished on an ordinary slide rule, considerable time and energy would be saved and the rule might then attain the same merited popularity with geologists that it has already won in engineering lines. At the same time its compact nature lends itself more readily to field use.

For these reasons the writer submits the accompanying tables (Tables I and II) of mineral percentage compositions, with a method of use applicable to any ordinary slide rule, which he trusts may meet the foregoing conditions successfully. In addition, conversion into molecular values by means of tables such as given by Osann[1] can be so much more quickly accomplished on a slide rule than by ordinary arithmetical processes.

	I*	II†
SiO_2	76.87	48.45
TiO_2	0.07	0.53
Al_2O_3	13.10	18.47
Fe_2O_3	1.05	4.81
FeO		3.96
MnO	0.01	0.06
MgO	0.06	3.50
CaO	0.45	7.38
Na_2O	3.55	4.58
K_2O	4.83	7.78
P_2O_5	0.01	0.48

* R. A. Daly, *Igneous Rocks and Their Origin*, 1914, p. 34, No. 97 (alaskite).
† *Ibid.*, p. 32, No. 90 (leucitite).

These tables contain two sets of minerals, one the more common of the rock-forming variety, the other the more important ore minerals. Precise mineral composition is, of course, only theoretically attained, but the formulae given are taken from standard texts[2] and represent generally accepted values. An

[1] A. Osann, *Beiträge zur Chemischen Petrographie*, I. Teil, "Molekular-Quotienten zur Berechnung von Gesteinsanalysen," Stuttgart, 1913.

[2] E. S. Dana, *Textbook of Mineralogy;* Moses and Parsons, *Mineralogy, Crystallography and Blowpipe Analysis;* F. W. Clarke, "Data of Geochemistry," *U.S. Geological Survey Bulletin 491;* Albert Johannsen, *Determination of Rock-forming Minerals in Thin Sections.*

average value for specific gravity is given instead of the range usually stated, as this average is accurate enough for such computations, and in fact is more accurate than the reconstructed analyses. This is due to the fact that the rocks of unlike mineral composition may have identical chemical composition, and also due to the range in chemical composition of such minerals as the biotites, amphiboles, pyroxenes, garnets, spinels, etc. The use of the tables may be explained most easily by taking two rock analyses and performing two conversions such as commonly confront the petrographer.

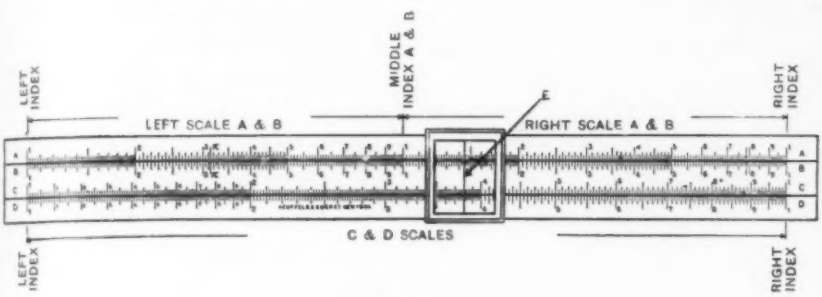

FIG. 1

The ordinary slide rule (see Fig. 1) has four graduated logarithmic scales, two on the immovable portion and two on the slide. For convenience these are referred to as "A," "B," "C," and "D," beginning with the upper one and reading downward. For brevity's sake the vertical line on the glass slide will be termed "E." A and B are duplicates, as are C and D. In using the mineralogical tables the scales C and D only are needed. First we shall proceed to reconstruct albite in Analysis I. The chemical analysis shows the presence of 3.55 per cent Na_2O. In Table I the percentage composition of albite is found to be

$$Na_2O—11.8; \quad Al_2O_3—19.5; \quad SiO_2—68.8$$

Set "E" on 118 on "D" and move "C" until 355 on this scale is under "E." Then if the corresponding amount of albite is desired.

move "E" to the left index on "D." The reading on "C" under "E" is 301, which means that 30.1 per cent of albite is equivalent to 3.55 per cent of Na_2O. Similarly, to find the amount of Al_2O_3 needed to combine with 3.55 per cent Na_2O in the albite molecule move "E" to 195 on "D." This will then cross "C" at 587, which indicates a requirement of 5.87 per cent of Al_2O_3. The sliding portion of the rule (carrying scales "B" and "C") has been set but once for both these readings. Now suppose that the SiO_2 value in albite corresponding to 3.55 per cent Na_2O is desired. In the table the SiO_2 percentage is 68.8. This value on "D" is not under "C" at all in the present setting. The next operation corresponds merely to doubling the length of the scale on "C." Place "E" on the right index of "C" and then move "C" to the right until the left index of "C" is under "E." Then move "E" to 688 on "D" and it is found to cross "C" at 20.7 which is the SiO_2 value required. A check on the correctness of these values thus obtained is to total the constituents and compare with the total obtained on "C."

Total on "C"	Separate Percentages
Albite—30.1 per cent	Na_2O—3.55 per cent
	Al_2O_3—5.87 " "
	SiO_2—20.7 " "
	Total—30.12 per cent

Now suppose we consider the potash in Analysis II which probably was present as leucite (7.78 per cent). Table I shows the composition of leucite as follows:

K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2

Set "E" on 215 on "D" and move the slide so that 778 on "C" is under "E," or is opposite 215 on "D." Move "E" to the left index on "D" and it is found to cross "C" at 36.19, which is the percentage of leucite present. In order to get the corresponding value of Al_2O_3 move "E" to 233 on "D." In this position it crosses "C" at 8.43, the percentage of Al_2O_3. The silica value, 55.2 on "D," again falls under the blank space in the rule, so set "E" on the right index of "C," move the slide to the right so that the left

index of "C" is under "E" and then set "E" on 552 on "D." In this position it crosses "C" at 19.98.

Total on "C"	Separate Percentages
Leucite—36.19 per cent	K_2O—7.78 per cent
	Al_2O_3—8.43 " "
	SiO_2—19.98 " "
	Total—36.19 per cent

The reverse operation, or conversion of minerals to their constituent oxides, will now be described. Suppose we take the case just cited. A rock contains 36.19 per cent leucite. What are the corresponding values of K_2O, Al_2O_3, and SiO_2? The percentage composition of leucite is

K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2

Set "E" over an index on "D," and since there is a choice of two such positions (one at each end of the scale) suppose the right index of "D" is chosen. Then move the slide so that the leucite percentage, 36.19 on "C," is under "E." It will be noted that of the three oxide values—21.5, 23.3, 55.2—only that of silica can be found on "D" under the slide in its present position. Hence set "E" on 552 on "D" and it will be found to cross "C" at 19.98, the percentage value of SiO_2. Then shift the slide so that 36.19 on "C" is opposite the left index on "D." Set "E" on 215 on "D" and it will cross "C" at 7.78, which is the potash value in the leucite percentage. Move "E" to 233 on "D" and it will cross "C" at 8.43, which is the Al_2O_3 value desired.

This explanation covers any ordinary case, including both the synthetical and the analytical processes. If the following point is borne in mind there should be no confusion. In the method as outlined, all percentage values from the table were read on "D," whereas all analytical values and their equivalents were read on "C." The opposite way is equally feasible, but if one way is always followed there is little excuse for confusion. Similarly, all these operations can be performed on scales "A" and "B," but the divisions are smaller and not so easily read.

TABLE I

Rock Minerals

Actinolite: 3.1—$CaMg_3FeSi_4O_{12}$: CaO—12.5; MgO—17.9; FeO—16.0; SiO_2—53.6
Aegirite: 3.55—$NaFeSi_2O_6$: Na_2O—13.4; Fe_2O_3—34.5; SiO_2—52.1
Albite: 2.63—$NaAlSi_3O_8$: Na_2O—11.8; Al_2O_3—19.4; SiO_2—68.8
Almandite: 4.05—$Fe_3Al_2Si_3O_{12}$: FeO—43.2; Al_2O_3—20.5; SiO_2—36.3
Alunite: 2.66—$KAl_3O_6H_6S_2O_8$: K_2O—11.4; Al_2O_3—37.0; SO_3—38.6; H_2O—13.0
Analcite: 2.25—$NaAlSi_2O_6H_2O$: Na_2O—14.0; Al_2O_3—23.2; SiO_2—54.6; H_2O—8.2
Andalusite: 3.18—Al_2SiO_5: Al_2O_3—62.9; SiO_2—37.1
Andesine: 2.68—Ab_3An_2: Na_2O—6.9; CaO—8.3; Al_2O_3—26.6; SiO_2—58.2
Andradite: 3.85—$Ca_3Fe_2Si_3O_{12}$: CaO—33.0; Fe_2O_3—31.5; SiO_2—35.5
Anhydrite: 2.94—$CaSO_4$: CaO—41.2; SO_3—58.8
Anorthite: 2.75—$CaAl_2Si_2O_8$: CaO—20.1; Al_2O_3—36.7; SiO_2—43.2
Apatite: 3.20—$Ca_4(CaF)P_3O_{12}$: CaO—55.5; P_2O_5—42.3; F—3.8
Augite: 3.31: CaO—20.9; MgO—12.6; FeO—6.9; Fe_2O_3—4.6; Al_2O_3—7.4; SiO_2—47.6
Barite: 4.45—$BaSO_4$: BaO—65.7; SO_3—34.3
Bauxite: 2.48—$Al_2O_3.2H_2O$: Al_2O_3—73.9; H_2O—26.1
Biotite: 2.95: K_2O—11.2; (Mg, Fe)O—19.2; Al_2O_3—24.4; SiO_2—43.1; H_2O—2.1
Calcite: 2.72—$CaCO_3$: CaO—56; CO_2—44
Chlorite (Penninite): 2.72: MgO—19.5; FeO—23.2; Al_2O_3—16.5; SiO_2—29.2; H_2O—11.6
Diopside: 3.33—$CaMgSi_2O_6$: CaO—25.8; MgO—18.6; SiO_2—55.6
Dolomite: 2.84—$CaMgC_2O_6$: CaO—30.4; MgO—21.7; CO_2—47.9
Enstatite: 3.17—$MgSiO_3$: MgO—40; SiO_2—60
Epidote (Pistacite): 3.38—$Ca_2(Al, Fe)_3HSi_3O_{13}$: CaO—23.5; Al_2O_3—24.1; Fe_2O_3—12.6; SiO_2—37.9; H_2O—1.9
Fayalite: 4.14—Fe_2SiO_4: FeO—70.6; SiO_2—29.4
Fluorite: 3.13—CaF_2: Ca—51.1; F—48.9
Forsterite: 3.25—Mg_2SiO_4: MgO—57.1; SiO_2—42.9
Gibbsite: 2.36—$Al(OH)_3$: Al_2O_3—65.4; H_2O—34.6
Grossularite: 3.57—$Ca_3Al_2Si_3O_{12}$: CaO—37.3; Al_2O_3—22.7; SiO_2—40.0
Gypsum: 2.32—$CaSO_4.2H_2O$: CaO—32.6; SO_3—46.6; H_2O—20.9
Halloysite: 2.1—$Al_2Si_2O_7.3H_2O$: Al_2O_3—36.9; SiO_2—43.5; H_2O—19.6
Hedenbergite: 3.6—$CaFeSi_2O_6$: CaO—22.2; FeO—29.4; SiO_2—48.4
Hornblende: 3.24: CaO—5.8; MgO—8.3; FeO—22.1; Fe_2O_3—16.3; Al_2O_3—10.5; SiO_2—37.0
Hypersthene: 3.45—(Mg, Fe)SiO_3: MgO—17.3; FeO—30.9; SiO_2—51.8
Kaolinite: 2.62—$Al_2Si_2O_7.2H_2O$: Al_2O_3—39.5; SiO_2—46.5; H_2O—14.0
Labradorite: 2.71—Ab_2An_3: Na_2O—4.6; CaO—12.3; Al_2O_3—30.0; SiO_2—53.1
Leucite: 2.48—$KAlSi_2O_6$: K_2O—21.5; Al_2O_3—23.3; SiO_2—55.2
Magnesite: 3.04—$MgCO_3$: MgO—47.6; CO_2—52.4
Melilite: 3.0: Na_2O—4.3; CaO—31.3; MgO—8.4; Fe_2O_3—11.2; Al_2O_3—7.1; SiO_2—37.7
Muscovite: 2.87—$H_2KAl_3Si_3O_{12}$: K_2O—11.8; Al_2O_3—38.5; SiO_2—45.2; H_2O—4.5
Nephelite: 2.6—$K_2Na_6Al_8Si_9O_{34}$: K_2O—7.7; Na_2O—15.1; Al_2O_3—33.2; SiO_2—44.0
Oligoclase: 2.66—Ab_3An_1: Na_2O—8.8; CaO—5.2; Al_2O_3—23.9; SiO_2—62.1
Orthoclase: 2.56—$KAlSi_3O_8$: K_2O—16.9; Al_2O_3—18.4; SiO_2—64.7
Serpentine: 2.56—$H_4Mg_3Si_2O_9$: MgO—43.0; SiO_2—44.1; H_2O—12.9
Staurolite: 3.70—$HFeAl_5Si_2O_{13}$: FeO—15.8; Al_2O_3—55.9; SiO_2—26.3; H_2O—2.0
Talc: 2.73—$H_2Mg_3Si_4O_{12}$: MgO—31.7; SiO_2—63.5; H_2O—4.8
Titanite: 3.52—$CaTiSiO_5$: CaO—28.6; TiO_2—40.8; SiO_2—51.7
Wollastonite: 2.85—$CaSiO_3$: CaO—48.3; SiO_2—51.7
Zircon: 4.69—$ZrSiO_4$: ZrO_2—67.2; SiO_2—32.8
Zoisite: 3.31—$HCa_2Al_3Si_3O_{13}$: CaO—24.6; Al_2O_3—33.7; SiO_2—39.7; H_2O—2.0

Oxides

Al_2O_3: Al—53.03; O—46.97
BaO: Ba—89.58; O—10.42
CaO: Ca—71.47; O—28.53
CO_2: C—27.27; O—72.73
Cr_2O_3: Cr—68.42; O—31.58
CuO: Cu—79.51; O—20.49
FeO: Fe—76.34; O—23.66
K_2O: K—83.02; O—16.98
MgO: Mg—60.32; O—39.68
MnO: Mn—77.44; O—22.56
Na_2O: Na—74.19; O—25.81
PbO: Pb—92.83; O—7.17
P_2O_5: P—43.69; O—56.31
SO_3: S—40.05; O—59.95

TABLE II

Ore Minerals

Iron

- Arsenopyrite: 6.05—$FeAsS_2$: Fe—34.3; As—46.0; S—19.7
- Chromite: 4.44—$FeCr_2O_4$: FeO—32; Cr_2O_3—68
- Goethite: 4.2—$FeO_3.H_2O$: Fe—62.9; O—27; H_2O—10.1
- Hematite: 5.1—Fe_2O_3: Fe—69.9; O—30.1
- Limonite: 3.8—$2Fe_2O_3.3H_2O$: Fe—59.8; O—25.7; H_2O—14.5
- Magnetite: 5.13—Fe_3O_4: Fe—72.4; O—27.6 (or FeO—31.0; Fe_2O_3—69.0)
- Pyrite: 5.03—FeS_2: Fe—46.6; S—53.4
- Pyrrhotite: 4.59—$Fe_{11}S_{12}$: Fe—61.5; S—38.5
- Siderite: 3.86—$FeCO_3$: FeO—62.1; CO_2—37.9
- Wolframite: 7.28—(Fe, Mn)WO_4: WO_3—76.5; (Fe, Mn)O—23.5

Copper

- Azurite: 3.8—$Cu_3(CO_3)_2(OH)_2$: CuO—69.2; CO_2—25.6; H_2O—5.2
- Bornite: 5.15—Cu_5FeS_4: Cu—63.3; Fe—11.2; S—25.5
- Chalcanthite: 2.21—$CuSO_4.5H_2O$: CuO—31.8; SO_3—32.1; H_2O—36.1
- Chalcocite: 5.65—Cu_2S: Cu—79.8; S—20.2
- Chalcopyrite: 4.2—$CuFeS_2$: Cu—34.5; Fe—30.5; S—35.0
- Chrysocolla: 2.13—$CuSiO_3.2H_2O$: CuO—45.2; SiO_2—34.3; H_2O—20.5
- Covellite: 4.59—CuS: Cu—66.5; S—33.5
- Cuprite: 6.0—Cu_2O: Cu—88.8: O—11.2
- Enargite: 4.44—Cu_3AsS_4: Cu—48.3; As—19.1; S—32.6
- Malachite: 3.97—$(CuOH)_2CO_3$: CuO—71.9; CO_2—19.9; H_2O—8.2
- Tennantite: 4.64—$Cu_8As_2S_7$: Cu—57.5; As—17.0; S—25.5
- Tetrahedrite: 4.8—$Cu_8Sb_2S_7$: Cu—52.1; Sb—24.8; S—23.1

Nickel and Cobalt

- Cobaltite: 6.07—CoAsS: Co—35.5; As—45.2; S—19.3
- Chloanthite: 6.5—$NiAs_2$: Ni—28.1; As—71.9
- Garnierite: 2.55—$H_2(Ni, Mg)SiO_4.xH_2O$: NiO—27.5; MgO—14.8; SiO_2—44.4; H_2O—13.3+
- Linnaeite: 4.9—Co_3S_4: Co—58; S—42
- Millerite: 5.53—NiS: Ni—64.7; S—35.3
- Niccolite: 7.49—NiAs: Ni—43.9; As—56.1
- Pentlandite: 4.75—(Fe, Ni)S_2: Ni—32.8; Fe—31.3; S—35.9
- Smaltite: 6.5—(Co, Ni)As_2: Co—22.0; Ni—21.9; As—56.1

Manganese

- Manganite: 4.3—$Mn_2O_3.H_2O$: Mn—62.4; O—27.3; H_2O—10.3
- Psilomelane: 4.2—H_4MnO_5: H—2.9; Mn—39.5; O—57.6
- Pyrolusite: 4.78—$MnO_2.nH_2O$: Mn—63.2; O—36.8
- Rhodochrosite: 3.5—$MnCO_3$: MnO—61.7; CO_2—38.3
- Rhodonite: 3.56—$MnSiO_3$: MnO—54.1; SiO_2—45.9

Silver

- Argentite: 7.29—Ag_2S: Ag—87.1; S—12.9
- Cerargyrite: 5.57—AgCl: Ag—75.3; Cl—24.7
- Dyscrasite: 9.64—Ag_3Sb: Ag—72.9; Sb—27.1
- Hessite: 8.41—Ag_2Te: Ag—62.9; Te—37.1
- Polybasite: 6.1—$(Ag, Cu)_9SbS_6$: Ag—75.6; Sb—9.4; S—15.0
- Proustite: 5.57—Ag_3AsS_3: Ag—65.4; As—15.2; S—19.4
- Pyrargyrite: 5.83—Ag_3SbS_3: Ag—59.9; Sb—22.3; S—17.8
- Stephanite: 6.25—Ag_5SbS_4: Ag—68.5; Sb—15.2; S—16.3

Lead

- Anglesite: 6.25—$PbSO_4$: PbO—73.6; SO_3—26.4
- Cerussite: 6.51—$PbCO_3$: PbO—83.5; CO_2—16.5
- Galena: 7.5—PbS: Pb—86.6; S—13.4
- Wulfenite: 6.85—$PbMoO_4$: PbO—60.7; MoO_3—39.3

Zinc

- Calamine: 3.45—$Zn_2SiO_4.H_2O$: ZnO—67.5; SiO_2—25.0; H_2O—7.5
- Smithsonite: 4.37—$ZnCO_3$: ZnO—64.8; CO_2—35.2
- Sphalerite: 4.02—ZnS: Zn—67.0; S—33.0
- Willemite: 4.04—Zn_2SiO_4: ZnO—73.0; SiO_2—27
- Zincite: 5.55—ZnO: Zn—80.3; O—19.7

Antimony Arsenic

- Orpiment: 3.48—As_2S_3: As—61.0; S—39.0
- Realgar: 3.52—As_2S_2: As—70.1; S—29.9
- Stibnite: 4.57—Sb_2S_3: Sb—71.4; S—28.6

Titanium

- Ilmenite: 4.7—$FeTiO_3$: Fe—36.8; Ti—31.6; O—31.6
- Rutile: 4.21—TiO_2: Ti—61.6; O—38.4 (Fe up to 10)

- Cassiterite: 6.95—SnO_2: Sn—78.6; O—21.4
- Cinnabar: 8.1—HgS: Hg—86.2; S—13.8
- Molybdenite: 4.75—MoS_2: Mo—60.0; S—40.0

As is true in many other instances, description is much slower and more tedious than manipulation, and the operations herein outlined may be performed easily in a few moments after one is accustomed to the use of the slide rule. To a person unfamiliar with the rule the question of placing the decimal point at once suggests itself. Used with the accompanying tables, the percentage values give all the assistance needed on this point and render the correct interpretation here more facile even than with the "geologist's slide rule." The values obtained should bear the same relations to each other as do the corresponding percentage values in the tables. For other operations the reader unaccustomed to slide-rule use is referred to various directions which are furnished by the manufacturers.

The accompanying tables are believed to be complete enough for most metamorphic, petrographic, and economic uses, but can be augmented to suit special needs. The percentages of a few oxides are included, as these are desirable in many cases. Any arrangement of minerals in such a table can be made to suit individual needs, and such tables mounted separately or back to back and covered with gelatin will be found an excellent form for field use, if such is desired.

LAKE SUPERIOR HIGHLANDS: THEIR ORIGIN AND AGE

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The dominant relief feature of the highland region about Lake Superior is an even, distinctly elevated plain in which the rivers are deeply intrenched. The genesis of this broad upland plain is a moot question of long standing. Its settlement involves the derivation and facial expression of the landscape over more than one-quarter of the entire North American continent.

At first glance the possibilities of the present peneplain's dating back in its formation to Early Cambrian, or even to pre-Cambrian, times, as is sometimes argued, seems so remote as almost to preclude serious consideration. Only by merest chance could there be survival of any remnant of so ancient a grade-plane. This chance lies in the exhuming of the old peneplain by the slow and uniform removal of a soft Paleozoic covering. Such a facet, if it persisted, would be quite small necessarily, coincidental, and, more properly, a product of some subsequent epoch of planation. Proofs of its antiquity have to be sustained by testimony overwhelmingly pertinent and convincing.

Indubitable evidence fixing the age of such a peneplain would not be likely to be found within the area of the highlands itself. It even might not be displayed anywhere within the limits of the great crystalline shield of Canada at all. Probabilities are for its disclosure far outside of the immediate elevated region. It is not a satisfactory solution of the problem to connect such an elevated and dissected plain with one of closely similar attitude emerging from beneath Cambrian sediments. In a continental geologic column there are many old plains—some only of provincial extent, but many of continental span or of dimensions as wide as such plains ever attain. To some place beyond the margins of the highland flats must attention first be turned for testimony bearing upon their age.

In casting about for a favorable locality in which to begin inquiry it is but natural to turn to some spot nearest to the highlands that displays a considerable section of the later geologic formations. There appears to be a section of this description off to the southwest of the middle angle of the Canadian shield, on the south side of the great tongue of pre-Cambrian rocks which extends from Lake Superior into northwestern Iowa where the old terranes are known as the Sioux quartzite.

On the general stratigraphic scheme of Iowa,[1] which is the latest and most complete of any now available, the erosion intervals may be readily indicated, together with the taxonomic values of each. Since some of these lines represent peneplanation sufficiently wide

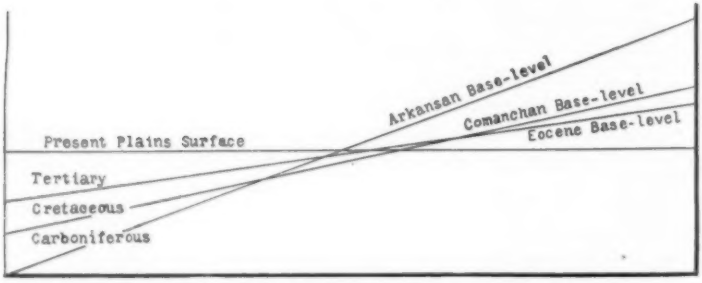

FIG. 1.—Relative attitudes of later base-levelings in Lake Superior region.

in extent to affect the Lake Superior region, they may be critically examined in turn, eliminated, or further compared in order that the parallelism with the highland plain under consideration may be exact.

Of the dozen or more conspicuous planes of unconformity which mark the geologic column of the region the Arkansan, Comanchean, and Eocene horizons are the most important. These represent most assuredly peneplains of great extent. Their intersections with the present ground-surface trend north-northwest through west-central Minnesota and Iowa. Along this line the intersections of all of them chance to be very close together. In diagram they may be represented as in Fig. 1.

[1] *Geol. Surv. Iowa*, XXII (1913), 154.

If the plane in which the lines of the diagram lie be considered as standing in a northeast-southwest direction the several peneplains of the Lake Superior region must have been successively obliterated, leaving only remnants of the very latest ones. The Arkansan peneplain, which constitutes the floor of the Coal Measures in the Upper Mississippi valley north of the Ozarks, may be for the present neglected. For reasons which are to follow, the Comanchean peneplain, which forms the Cretaceous floor in Iowa, Minnesota, and Manitoba, may be particularly examined.

Recent investigations about the point where the three states of Iowa, Minnesota, and South Dakota meet clearly disclose some instructive geologic structures bearing directly upon the problem under consideration. The field data thus acquired are supplemented by numerous deep-well records. By reference to the general geologic map of Iowa the Paleozoic formations are noticed to be distributed in relatively narrow belts trending in a northwest direction across the northeastern one-third of the state. Very singularly, it has always seemed, these belts abruptly terminate at the north soon after the state boundary is passed. This rather peculiar circumstance appears never to have excited curiosity as to its cause. Far to the north, about Winnipeg, in Manitoba, there is the same narrow belting of the same formations and, as farther south, the strike is northwest. The Canadian Paleozoic area is separated in central Minnesota from the Iowan Paleozoic field by a broad expanse of pre-Cambrian rocks.

Structurally these pre-Cambrian rocks form the core of a rather notable arch the axis of which runs northeast and southwest. This anticline is one of large proportions and extends from the east shore of Lake Superior to central South Dakota, where, as a canoe-shaped form, it plunges beneath the post-Paleozoic deposits of the Great Plains. The exposure of Sioux quartzite constitutes the western nose of the fold.

It is against the south slope of the sharp Siouan anticline that the belted Paleozoic terranes of northeastern Iowa are upturned, and there cut off. The eastern margin of the vast Cretaceous field crosses the same line, so that there is apparently no westward extension of the five groups of formations, if it ever existed, at

least on the surface of the ground. On the other, or north, side of the anticline the same belts recur, as already stated.

Bearing in mind the geographic position of this marked anticline, an arch between the center of which and the bases of its limbs there is a stratigraphic interval of more than 5,000 feet, it is quite obvious that the Paleozoic belts originally did not really terminate against it in southern Minnesota but rather extended over it and were continuous with the similar Canadian belts. This being the case, it is equally obvious that the Iowan belts should not only not stop against the arch, but should continue westward along the strike of the fold, but beneath the Cretaceous covering. This is found actually to accord with recently observed facts. A cross-section (Fig. 2), which is drawn to scale, indicates the actual amount of tilting displayed at the present time, with the part originally laid down, but removed during Mid-Cretaceous times, represented by dotted lines.

There are incontrovertible proofs fixing within very narrow limits the geologic date of the uprising of the great Siouan arch, and also the time of its complete reduction again to an even plain lying but little above the level of the sea. Since all of the Paleozoic formations, from Cambrian to latest Carboniferous, take part in the folding while the Cretaceous strata do not, it is manifest that the main movement occurred in Early Mesozoic times, largely during the Triassic and Jurassic periods. Comanchean time (Early Cretaceous) in the region must have been principally a period of rapid, enormous, and very complete denudation, since by the beginning of Mid Cretaceous time, when marine deposition over this part of the continent took place, the floor upon which the sediments were laid down was as even as any known peneplain. The Cretaceous floor is a true plain worn out on the beveled edges of the Paleozoics and older rocks.

The areal distribution and attitude of the Cretaceous beds along their eastern margin is suggestive. In Minnesota the western half of the state is occupied by deposits of Cretaceous age. Outliers occur far to the eastward—to the Mississippi River and the Mesabi Range. The great thickness of Cretaceous deposits in the eastward-facing escarpment of the Duck and Riding mountains, in

Manitoba, and their gentle dip to the west point to their former great eastward extension probably to Lake Superior and Hudson Bay. There is little reason to doubt, therefore, the assumption that the Cretaceous sediments once covered the Lake Superior

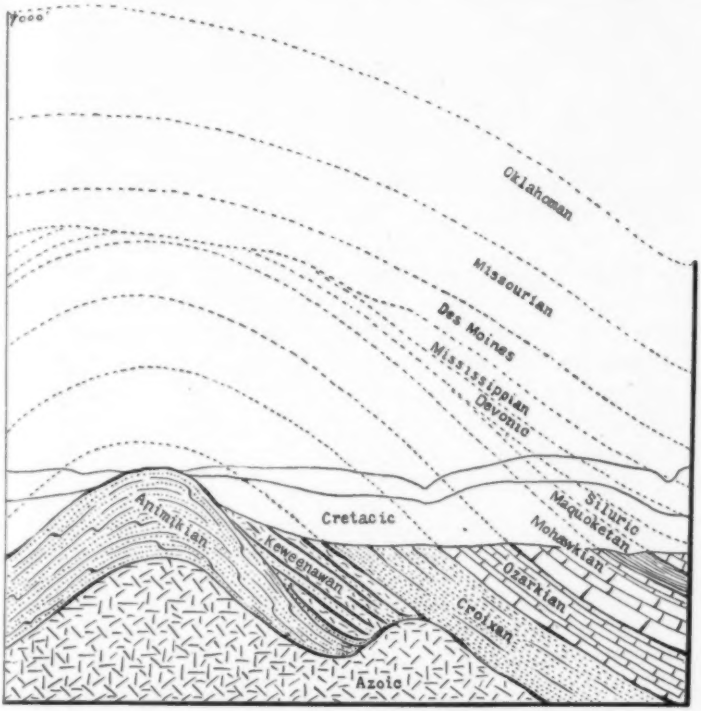

FIG. 2.—Erosion of the Triassic Siouan mountains in southern Minnesota.

highlands and that the peneplain which formed the Cretaceous floor must have imparted the main topographic expression to the highland plain.

In Iowa, and Minnesota also, the Cretaceous strata themselves are notably beveled, but at a lower angle than that of their floor. There are good reasons for believing that this Tertiary base-leveling took place in Eocene time. There appear to be important water-

laid deposits of this age in the region; but Miocene and Pliocene deposits seem to be chiefly, if not entirely, continental in character.

It is quite probable, therefore, that the tops of the monadnocks which rise above the level of the highland plain of Lake Superior represent approximately the level of the Comanchean peneplain; and that the present plain of that region is Eocene in date. This interpretation has many points supporting it; and but few or none against it. Around Lake Superior glacial action has entirely removed the evidence for the ready determination of these supporting facts. The strongest and most convincing testimony comes from neighboring localities; and the several lines of evidence appear to be wholly congruous and mutually supporting.

For the formation of the Lake Superior highlands an Eocene age appears to be conclusive. The necessary consequences are far-reaching. It calls for an examination anew of the geographic development of a large part of the northeastern section of the North American continent.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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V. QUEBEC

Barlow[1] states that the Chibougaman region of Quebec is underlaid by Keewatin, Laurentian, and Huronian formations, and a gabbro anorthosite.

Barlow[2] states that the Chibougaman region, lying between latitudes 49° 30′ and 50° 11′ 06″ north and longitude 70° 03′ 40″ to 74° 35′ 50″, about 300 miles north of Montreal, has the following pre-Cambrian succession region, beginning with the oldest: (1) Keewatin, serpentine and pyroxenite; deformed quartz porphyry and porphyrites, basalts, diabases and gabbros; (2) anorthosite intrusive into Keewatin; (3) Laurentian granites, granite and diorite gneiss intrusive into (1) and (2); (4) unconformity; (5) Lower Huronian—slates, sandstones, arkose and conglomerate.

Bancroft[3] classifies the pre-Cambrian rocks of the Harricanaw and Nottaway rivers of northwestern Quebec as Keewatin metamorphosed extrusive and intrusive basic and acid igneous rocks with bands of highly altered sedimentary rocks; Laurentian granites and granite gneisses intrusive into the Keewatin; Lower Huronian (?) conglomerates, graywacke, slate, and arkose, injected by granite; post-Middle Huronian (Keweenawan ?) dikes of quartz and olivine diabase.

[1] "Geology of the Chibougaman Region, Quebec," *Canada Geol. Soc. Am. Bull.*, XXII, No. 4 (1911), 738 pp.

[2] "Geology and Mineral Resources of the Chibougaman Region," Report by the Chibougaman Mining Concession, *Province of Quebec, Canada*, 1911, 215 pp., 68 pls., 19 figs., 2 maps.

[3] J. A. Bancroft, "Geology and Natural Resources of the Basins of Harricanaw and Nottaway Rivers, Northwestern Quebec," *Province of Quebec*, 1912, 16 pp., 1 map.

Cirkel[1] states the Amherst graphite deposits occur as lenticular masses and pockets, disseminations, and veins in eruptive rock intrusive into the Grenville limestones. Graphite occurs within feldspar, quartz, and pyroxene. It is mostly associated with feldspar or pyroxene, less frequently with calcite and wollastonite. Garnet, zircon, scapolite, titanite, muscovite, pyrite, apatite, leucoxene, biotite and monazite are accessory minerals of the gangue.

Dulieux[2] reports that the rocks of the Lake Chibŏgomon region, north of Lake St. John, in Quebec, constitute an isolated belt of diabases and gabbros and rocks derived from them; viz., chloritic, epidotic, talc schists, and serpentine, surrounded by Laurentian granites and gneisses, the backbone of Labrador. They contain deposits of cupriferous and auriferous sulphides, magnetic iron, and asbestos.

Wilson[3] finds that the pre-Cambrian rocks in western Quebec between 75° 30′ and 79° 30′ west longitude and 48° and 49° north latitude include hornblende schists, chlorite schists, diabase, and porphyries, all probably of Keewatin age, and Laurentian granites and gneisses.

VI. THE CORDILLERAS

Ball[4] finds that the pre-Cambrian, magnetic, titaniferous iron ores of Iron Mountain in southeastern Wyoming form a series of dikes cutting anorthosite. Granitic masses dissect both. The ores and the anorthosite contain the same minerals with some minor exception, but in different proportions. The tonnage of ore is large, averaging high in iron, and probably low in phosphorus, but too high in titanium to be desirable under present conditions.

[1] Fritz Cirkel, "The Amherst (Quebec) Graphite Deposits," *Quart. Bull. Can. Mining Inst.*, 1910, pp. 107–15.

[2] E. Dulieux, "The Chibŏgomon Region, Quebec," *Jour. Can. Min. Inst.*, XII (1909), 184–93.

[3] W. J. Wilson, "Geological Reconnaissance along the Line of the National Transcontinental Railway in Western Canada," *Canada Geol. Survey, Mem.* 4, (1910), pp. 56, 5 pls., 1 geol. map.

[4] Sidney H. Ball, "Titaniferous Iron Ore of Iron Mountain, Wyoming," *Bull. 315*, 1906, pp. 206–12.

Bancroft[1] states that the oldest rocks in Northern Yuma County, Arizona, are granites and diorites, probably of Archaean age. They are capped by a thick series of metamorphosed pre-Cambrian sediments, quartzites, limestones, dolomites, argillites, arenaceous shales, and quartz mica schists, interlayered and intruded by basic rocks altered to amphibolites.

Blackwelder[2] reports that the pre-Cambrian of three geologic sections in western Wyoming consists of gneiss, gray, and red granites penetrated by diabase and other dikes, hornblende gneiss, and granite with large dikes of diabase.

Blackwelder[3] states that the pre-Cambrian rocks of the Laramie and Sherman quadrangles of northeastern Wyoming include: (1) hornblende schists, associated with some schistose rhyolites and felsites, soft mica gneisses et cetera, and rocks resembling contorted quartzites, and limestones, all highly metamorphosed; these rocks are the oldest in the area; (2) mildly metamorphosed granite gneiss representing two periods of intrusion, both younger than (1); (3) basic intrusives, comprising syenites, gabbros, diorites, granodiorites, and gabbro gneiss, all younger than (2); (4) granite porphyry intrusive into (1) and (2); (5) anorthosite, intrusive into (1), (2), and the gabbro gneiss of (3); (6) granite intrusive into (1), (2), (3), (4), and (5); (7) small diabasic and dioritic dikes injected into (6) and probably younger than (7).

Blackwelder[4] finds an unconformity in Big Cottonwood Canyon along the upper course of the Ogden River in the Wasatch Mountains, between slightly fossiliferous early Cambrian quartzites and an older quartzite slate series, several thousand feet thick, which he assigns to the Algonkian in accordance with the prevailing methods of correlation.

[1] Howland Bancroft, "Reconnaissance of the Ore Deposits in Northern Yuma County, Arizona," *Bull. 451, U.S. Geol. Survey*, 1911, pp. 130, maps and illustrations.

[2] Eliot Blackwelder, "A Reconnaissance of the Phosphate Deposits in Western Wyoming," *Bull. 470, U.S. Geol. Survey*, 1911, pp. 452–81.

[3] N. H. Darton, E. Blackwelder, C. E. Siebenthal, "The Laramie and Sherman Quadrangles, Wyoming," *U.S. Geol. Survey, Folio No. 173*, 17 pp., 8 pls., 3 figs.

[4] Eliot Blackwelder, "New Light on the Geology of the Wasatch Mountains, Utah," *Bull. Geol. Soc. Am.*, XXI (1910), 517–42, pls. 36–40.

The unconformity is shown by the variable thickness of the quartzite-slate series, slight discordance of bedding, and a basal conglomerate.

The Algonkian system consists of alternating quartzites, slates, and conglomerates, which are variable from place to place; and show cross-bedding, ripple marks, and mud cracks. The materials are not well assorted, and in the sandy beds, the prevailing colors are shades of red; green predominates in the shales. Fossils and limestone appear to be lacking. Blackwelder is inclined to view these sediments as subaerial, largely fluviatile deposits.

Cross and Hole[1] state that the pre-Cambrian in the Engineer Mountain quadrangle of Colorado include Archaean gneisses and schists derived from granites, diorites, and diabase. These are intruded by granitic and gabbroic masses. Several thousand feet of massive and some thin-bedded quartzites, white, brown, and black, and occasional bands of brown to black shale, all of Algonkian age, overlie the Archaean.

Diller[2] describes the occurrence of asbestos in the pre-Cambrian rocks near Casper, Wyo., and the Grand Canyon of Arizona. At Casper, Wyo., the asbestos occurs mostly at the igneous contact between serpentine and granite. In the Grand Canyon, asbestos occurs only where diabase sills have been intruded into Algonkian magnesian limestones.

Diller[3] states that the rocks of the asbestos area south and southeast of Casper, Wyo., are hornblende schist, diorite, granite, and serpentine of pre-Cambrian age.

In the Grand Canyon of Arizona, certain Algonkian limestone beds contain asbestos where they have been intruded by diabase sills.

Daly[4] finds that the great stratigraphic problem of the Shuswap Lakes region is the correlation of the non-fossiliferous, crystalline

[1] W. Cross and A. H. Hole, "Engineer Mountain Folio," *U.S.G.S. Folio* 171, 1910, 14 pp., 5 pls., 1 fig.

[2] J. S. Diller, "The Types and Modes of Occurrence of Asbestos in the United States," *Jour. Can. Min. Inst.*, 1911, pp. 93–107.

[3] J. S. Diller, "The Types, Modes of Occurrence, and Important Deposits of Asbestos in the United States," *Bull. 470, U.S. Geological Survey*, 1911, pp. 505–23.

[4] R. A. Daly, "Reconnaissance of the Shuswap Lakes and Vicinity (South Central British Columbia), *Canada Department of Mines*, Ottawa, 1912.

rocks which lie unconformably beneath the Carboniferous. Dawson found that an unconformity separated them into two divisions. The rocks beneath this unconformity he classed as pre-Cambrian and those above as Cambrian. Daly classifies all the rocks unconformably beneath the Carboniferous as pre-Cambrian, excepting a part of the quartzite rocks immediately underlying the Carboniferous which he believes may be Cambrian. The unconformity which Dawson believed separated the Cambrian from the pre-Cambrian Daly places within the pre-Cambrian, and calls the rocks below this unconformity pre-Beltian, and those above it Beltian, with the reservation previously stated that a part of the upper portion may be Cambrian.

The pre-Beltian consists of the dominantly Sedimentary Shuswap series, about 26,500 feet thick, which is intruded by batholiths, dikes, and sills of granite. It includes limestones, metargillites, mica-schists, paragneisses, green schists and greenstones, the latter probably of extrusive origin although no definite extrusive characteristics have been found. The Beltian is also dominantly sedimentary, consisting from the base up of about 200 feet of arkose sandstone overlain by about 18,000 feet of argillaceous rocks interbedded with limestones and quartzites, the whole constituting the Albert Canyon division. Above this is the Glacier quartzite division about 25,000 feet thick, the upper part of which, Daly believes, may be Cambrian, although no fossils have been found.

The strike of the pre-Cambrian rocks is nearly east and west or practically at right angles to the Cordilleran folding. Normal faulting in part contemporaneous with deposition is more evident than folding, and the bedding is either nearly horizontal or only slightly inclined. From the fact that both the Shuswap and the Beltian are dominantly schistose with their schistosity essentially parallel with the slightly deformed bedding. Daly infers that the schistosity has not resulted from tangential stress but from the weight of the sediments and the action of thermal solutions. While the pre-Cambrian rocks are only gently inclined, the Carboniferous and Triassic rocks of this region are highly folded.

The rocks of the Rocky Mountain geosynclinal are believed by Daly to be largely derived from the pre-Beltian. He finds that the microscopic characteristics of the feldspars in the sediments are identical with those of the granites, sills, and dikes which intrude the Shuswap series. The extraordinary fine grain of the quartzites he attributes to their derivations from metargillites, phyllites, and greenstone schists.

Daly does not care to have the term Algonkian applied to the pre-Cambrian of the Shuswap Lakes region, since it would include two series which are separated by a well marked unconformity.

Darton[1] records a few observations on the distribution of the Vishnu and Grand Canyon series in western New Mexico and northern Arizona.

Emmons[2] states that a complex of pre-Cambrian granites constitutes the basement upon which later sedimentary and igneous rocks of the Cananea district were superimposed.

Emmons[3] states that the oldest rocks of the Little Rocky Mountains between the Missouri and Milk rivers consist of pre-Cambrian quartzite by intrusive porphyry.

George and Crawford[4] find that the pre-Cambrian rocks of the Hahn's Peak region of Routt County, Colo., as a complex of schists, gneiss, gneissoid granite, and granite with isolated patches of quartzite.

Henderson[5] describes the pre-Cambrian rocks of the Front Range of Colorado, as consisting of Archaean granite and gneiss, and an Algonkian quartzite, outcropping from South Boulder Canyon

[1] N. H. Darton, "A Reconnaissance of Parts of Western New Mexico and Northern Arizona," *Bull 435, U.S. Geological Survey*, 1910, pp. 14–15.

[2] S. F. Emmons, "Cananea Mining District of Sonora," *Econ. Geol.*, V, No. 4 (1910), 312–.

[3] W. H. Emmons, "Gold Deposits of the Little Rocky Mountains, Montana," *Bull. 340, U.S. Geological Survey*, 1907, pp. 96–116.

[4] R. D. George and R. D. Crawford, "The Hahn's Peak Region, Routt County, Colorado," *Colo., Geol. Surv., 1st Rept.*, 1908, pp. 189–229, 1 pl., 1909.

[5] Junius Henderson, "The Foothills Formations of Northern Colorado," *Colo. Geol. Surv., 1st Rept.*, pp. 149–85, 1909.

southward a distance of about twelve miles. The latter is mostly white and in places coarsely conglomeratic, showing both bedding and cross-bedding.

Lindgren, Groton, and Gordon[1] find small areas of pre-Cambrian rocks, most of them in the northern part of New Mexico, exposed in eroded regions of doming and faulting, flanked by younger rocks. The largest, about 120 miles long, extends as a southward continuation of the Sangre de Cristo range of Colorado, to about 20 miles south of Santa Fe. It consists of a sedimentary series, chiefly quartzitic, which with certain greenstone tuffs, amphibolites, and rhyolites has been invaded by powerful intrusions of normal, usually reddish microcline granite, locally schistose. The granite in turn has been intruded by aplite, abundant pegmatite, and some masses or dikes of greenstone. The sediments are believed to correspond to those of Colorado similarly imbedded in red granite and to the quartzitic schists of the Pinal formation of Arizona, but the pre-Cambrian sediments of the Grand Canyon are believed to be younger than the red granite intrusions.

Unimportant ore deposits of gold and silver, some zinc and copper, but little lead are found in the granites, gneisses, and amphibolitic schists. Some rich gold placers have developed from them. The ores occur in quartz-filled fissures, usually of the "lenticular" type, shear zones filled with quartz stringers, and disseminations of sulphides in amphibolitic schists. The deposits frequently show evidence of deep-seated origin, such as the development of heavy minerals, and massive, glassy vein quartz. Many of them were folded and fractured. The ore deposition followed granitic and other intrusions, and the metals were probably delivered from the magmas.

Noble[2] describes the pre-Cambrian succession of the Shimuno area of the Grand Canyon. The following is a summarized section:

[1] W. Lindgren, L. C. Groton, and C. H. Gordon, "The Ore Deposits of New Mexico," *Prof. Paper 68, U.S. Geological Survey*, 1910, pp. —, several maps.

[2] L. F. Noble, "Contributions to the Geology of the Grand Canyon, Arizona. The Geology of the Shimuno Area, *Am. Jour. Science*, 4th Series, XXIX, 369–86.

	UNCONFORMITY AT BASE OF CAMBRIAN
	Micaceous shaly sandstone uniform in composition but varying in color. Ripple marked and cross-bedded throughout. 2,297 feet thick.
	Basal conglomerate, 6 feet
Unkar Group.	Calcareous shales and dolomitic limestones and dolomites containing some ripple marks, sun-cracks; total thickness, 304 feet.
Thickness, 478*a* feet.	Ripple-marked, cross-bedded jaspers, quartzitic jasper, intrusive diabase, blue, red, vermilion, argillaceous and arenaceous shales. Thickness 579 feet.
	Pure fine-grained sandstones and quartzites, ripple-marked and cross-bedded. 1,564 feet thick.
Unconformity.	
Vishnu series.	Quartz schist, quartz mica schist, quartz hornblende schist, hornblende schist, quartz diorite and pegmatite granite.

Paige[1] states that the pre-Cambrian complex of Grant County, New Mexico, comprises various types of coarse-to-medium-grained granites and porphyries, and pegmatitic dikes.

Patton[2] finds that the pre-Cambrian rocks of the Grayback Mining district of Costilla Co., Colorado, consist of granite, granite and biotite gneiss, and a small amount of hornblende schist, amphibolite, and pegmatite.

Pardee[3] describes the sedimentary rocks of the Upper St. Joe River in Idaho which are supposed to be equivalent to the Algonkian formations of the Coeur d'Alene district. They consist of sandstones, graywackes, quartzites, shales, garnetiferous mica schists, and limestone.

Ransome[4] states that the basal rocks of the Breckenridge district

[1] Sidney Paige, "Metalliferous Ore Deposits near the Burro Mountains, Grant County, New Mexico," *Bull. 470*, 1911, *U.S. Geological Survey* (1911), pp. 131–50, 1 map.

[2] H. B. Patton et al., "Geology of the Grayback Mining District," *Colo. Geological Survey, Bull., 2*, 1910, 111 pp., 9 pls.

[3] J. T. Pardee, "Geology and Mineralization of the Upper St. Joe River, Idaho," *Bull. 470*, 1910, pp. 39–61, 1 map.

[4] F. L. Ransome, "Geology and Ore Deposits of the Breckenridge District," *U.S.G.S., Prof. Paper 75*, 1911, pp. 187.

of Colorado appear as isolated patches of pre-Cambrian schists and gneisses cut by pegmatite and granite.

Schrader[1] states that the principal topographic features of Mohave County, Arizona, are barren, plainlike, detritus-filled valleys 10 or 12 miles in width, separated by ranges of about equal width. Their trend is a little west of north, and the general slope is toward the southwest.

The ridges consist mainly of pre-Cambrian granites, porphyries, gneisses and schists cut by pre-Tertiary intrusives. The pre-Cambrian probably was once covered or nearly covered by Paleozoic sediments which are now almost completely eroded.

Schrader[2] finds that the pre-Cambrian granites, gneisses, and schists constitute the basement complex on which later sediments of Mohave County, Arizona, have been deposited. No pre-Cambrian sediments have been recognized. The trend of the schists is about N. 30° E. and their dip is either vertical or steeply east-northeast.

Smith[3] believes that the older gneisses of the Sierra Nevada are probably Archaean.

Spurr[4] states that the lowest formation in the Aspen district is pre-Cambrian granite.

Stone[5] finds that in the southeastern part of T. 5 N., R. 2 W., of Montana there is a flat area, underlain by Algonkian sediments of unknown thickness, mostly red shales, including some green shale and thin sandstone beds. They are believed to correspond to the Spokane shale to the east.

[1] F. C. Schrader, "The Mineral Deposits of Mohave County, Arizona," *Bull. 340, U.S. Geological Survey*, 1907, pp. 53–95.

[2] F. C. Schrader, "Mineral Deposits of the Cerbat Range, Black Mountains, and Grand Wash Cliffs, Mohave County, Arizona," *Bull. 397, U.S. Geological Survey*, 1909, pp. 220.

[3] James Perrin Smith, "The Geological Record of California," *Jour. Geol.*, XVIII, No. 3 (1910), 216–27.

[4] J. Edward Spurr, "Ore Deposition at Aspen, Colorado," *Econ. Geol.*, IV (1909), 301–20.

[5] R. W. Stone, "Geologic Relations of Ore Deposits in the Elkhorn Mountains, Montana," *Bull. 470, U.S. Geological Survey*, 1911, pp. 75–98, 1 map.

Turner[1] describes the complex of pre-Cambrian granite-gneiss, quartz-monzonite gneiss, granite augen schists, calcareous augen schists, and small lenses of hydrous mica-schists of the Silver Peak district of Nevada. Certain dolomites, quartzites, and green knotted schists underlying the Olenellus fauna may be Algonkian, but have been provisionally referred to the Lower Cambrian.

Walcott[2] finds the following succession of pre-Cambrian sediments exposed below basal Cambrian conglomerate in the Bow River valley of Alberta, Canada: Hector formation—shales 1,300–2,150 feet; Corral Creek formation—sandstones 2,600 feet, base not found. He correlates them with the Camp Creek and Sheppard series of arenaceous rocks which lie beneath the Cambrian and above the Siyet limestone in Montana, southwestern Alberta, and southeastern British Columbia.

Weeks[3] finds that the oldest rocks of the Osceola district are granites and schists, with intruded porphyry, which he believes are probably Archaean.

Winchell[4] finds that the lowest rocks exposed in the Ruby range, about 60 miles south of Butte, Mont., are quartz schists and slates, probably of pre-Cambrian age.

[1] H. W. Turner, "Geology of the Silver Peak Quadrangle," *Bull. of Geol. Soc. Am.*, XX (1909), pp. 223–64, 4 pls., 1 fig.

[2] C. D. Walcott, "Pre-Cambrian Rocks of the Bow River Valley, Alberta, Canada," *Smith. Misc. Coll.*, LIII, No. 7 (August, 1910), pp. 423–31, 3 pls.

[3] F. B. Weeks, "Geology and Mineral Resources of the Osceola Mining District, White Pine County, Nevada," *Bull. 340, U.S. Geological Survey*, 1907, pp. 117–33.

[4] Alexander N. Winchell, "Graphite near Dillon, Montana," *Bull. 470, U.S. Geological Survey*, 1911, pp. 528–32.